HYDROLOGIC ISSUES IN ARID, UNSATURATED SYSTEMS AND IMPLICATIONS FOR CONTAMINANT TRANSPORT

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Abstract. Analysis of unsaturated flow and transport in arid regions is important, not only in water resource evaluation but in contaminant transport as well, particularly in siting waste disposal facilities and in remediating contaminated sites. The water fluxes under consideration have a magnitude close to the errors inherent in measuring or in calculating these water fluxes, which makes it difficult to resolve basic issues such as direction and rate of water movement and controls on unsaturated flow. The purpose of this paper is to review these issues on the basis of unsaturated zone studies in arid settings. Because individual techniques for estimating water fluxes in the unsaturated zone have limitations, a variety of physical measurements and environmental tracers should be used to provide multiple, independent lines of evidence to quantify flow and transport in arid regions. The direction and rate of water flow are affected not only by hydraulic head gradients but also by temperature and air pressure gradients. The similarity of water fluxes in a variety of settings in the southwestern United States indicates that vegetative cover may be one of the primary controls on the magnitude of water flow in the unsaturated zone; however, our understanding of the role of plants is limited. Most unsaturated flow in arid systems is focused beneath topographic depressions, and diffuse flow is limited. Thick unsaturated sections and low water fluxes typical of many arid regions result in preservation of paleoclimatic variations in water flux and suggest that deep vadose zones may be out of equilibrium with current climate. Whereas water movement along preferred pathways is common in humid sites, field studies that demonstrate preferential flow are restricted mostly to fractured rocks and root zones in arid regions. Results of field studies of preferential flow in humid sites, generally restricted to the upper 1–2 m because of shallow water tables, cannot be applied readily to thick vadose zones in arid regions.

1. INTRODUCTION

In the past, unsaturated-zone studies in arid settings were conducted primarily for water resource evaluation. During the past 2 decades, however, emphasis has shifted from water resources to waste disposal and contaminant transport. In addition to remediation of contaminated sites in arid regions, arid areas are also being proposed for low-level and high-level radioactive waste disposal [*Montazer and Wilson*, 1984; *Scanlon*, 1992a; *Prudic*, 1994]. Water resource evaluation studies generally assume uniform rates of water movement throughout a study area because that assumption may not greatly affect resource estimates. In contrast, application of uniform rates of water movement to contaminant transport analyses in areas of spatially variable water movement could invalidate estimated rates of contaminant transport. Knowledge of spatial variability in unsaturated flow is therefore critical for realistic assessment of transport rates because such spatially variable rates could allow contaminants in some areas to migrate rapidly, essentially bypassing the buffering capacity of much of the unsaturated zone.

Low precipitation rates and high evapotranspiration rates should result in low rates of water movement in arid settings. The book *Deserts as Dumps* by *Reith and Thomson* [1992] evaluates many issues related to waste disposal in arid regions. Groundwater contamination in many arid settings such as Hanford, Washington [*Dresel et al.*, 1996], Sandia, New Mexico [*Crowson et al.*, 1993], and the Negev Desert, Israel [*Nativ et al.*, 1995], has resulted in considerable debate about the suitability of arid settings for waste disposal. In the past, National Academy of Science (NAS) panels suggested that arid sites are unsuitable for radioactive waste disposal because of limited information on flow processes in arid regions [*National Research Council* (*NRC*), 1957, 1966]. The findings of a recent NAS panel suggest, however, that interstream settings in arid regions should be suit-

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able for waste disposal [*NRC*, 1995]. Does this shift in opinion reflect an increased understanding of unsaturated flow and transport processes in arid settings?

Much research on unsaturated-zone hydrology has been conducted in humid sites; however, fundamental differences between humid and arid regions limit the applicability of techniques developed at humid sites to arid sites. Such fundamental differences include thickness of the unsaturated zone, which can be as much as several hundred meters in arid regions but commonly is only meters thick in humid sites. Water fluxes and water content of unsaturated media also have a much greater range in arid sites than in humid sites. Greater thickness of the unsaturated zone and lower water fluxes in many arid settings result in much longer timescales being represented by unsaturated sections in arid regions (up to thousands of years) than in humid regions (up to tens of years). Because of these differences the results of studies conducted in humid regions should not be applied directly to arid regions.

Questions about the suitability of arid settings for waste disposal may result from limited understanding of unsaturated-flow processes, in turn reflecting the limitations of various techniques for quantifying the extremely low water fluxes typical of interfluvial settings in many arid regions. As a result of low water fluxes and the limitations of various techniques to quantify flow, basic issues such as (1) direction and rate of water movement and (2) mechanisms and controls of water flow are difficult to resolve. The purpose of this paper is to examine some of the basic issues related to unsaturated flow by reviewing unsaturated-zone studies in arid settings. Some of the issues that will be considered are as follows:

1. What are the difficulties inherent in various techniques used to evaluate flow and transport?
2. What are the direction and rate of water movement?
3. How important is preferential flow in arid regions?
4. What are the most important controls on water flow and transport?
5. What is the role of vegetation in controlling water flow?
6. What effect do potential climate changes have on unsaturated flow?
7. How can we numerically simulate flow in arid settings?

An understanding of these issues is important for evaluation of water resources in arid regions and also for analysis of contaminant transport related to municipal, hazardous, and radioactive waste disposal.

Although arid regions occur throughout the world, unsaturated-zone studies have been conducted primarily in the western United States and in Australia; limited studies have been conducted in Africa, Israel, and Saudi Arabia. Results of studies of these arid settings are evaluated in this paper to provide insights into some of the basic issues described above.

Most of the studies referenced in this paper were conducted in the western United States. These studies include remediation of contaminated areas such as at Hanford, Washington [*Dresel et al.*, 1996] and Sandia (near Albuquerque), New Mexico [*Crowson et al.*, 1993], and at several uranium mill tailings sites [*Reith and Thomson*, 1992]. In addition, arid sites have been proposed for low-level radioactive waste disposal (from medical and research activities, and power plants) in Ward Valley, California, and Eagle Flat, Texas. Commercial facilities for disposing of low-level radioactive waste include Richland, Washington, and Beatty, Nevada (1962–1992). Deep (~300 m) geological disposal in the unsaturated zone at Yucca Mountain, Nevada, is proposed for high-level radioactive waste, which includes spent fuel from nuclear power plants and material from the nuclear weapons industry. Because much of the waste remains radioactive for a long time, we are concerned not only with flow and transport in the natural system, which can serve as a long-term (hundreds to thousands of years) barrier, but also with how we can engineer systems so as to minimize water fluxes.

To evaluate flow processes in the unsaturated zone, we need detailed information at small scales (~0.3 m); however, results from small-scale studies may have implications for much larger areas. Timescales of interest range from days to thousands of years, depending on the problem being evaluated. Arid systems are generally characterized by episodic flow that can occur in days in response to a sequence of precipitation events. In contrast, the period of time required for high-level nuclear waste to remain isolated from the accessible environment is ~10,000 years [*NRC*, 1995].

First we evaluate various techniques for quantifying unsaturated flow that use both hydraulic and hydrochemical approaches. Then we discuss the various driving forces for water movement that control the direction of water flow. Next we review preferential flow and how important it is in desert systems. The controls on unsaturated flow, including vegetation, climate, texture, and topography, are evaluated with reference to published studies. Recent improvements in numerical modeling that apply to simulations of flow and transport in arid regions are discussed, and results of case studies are presented. We close the discussion with some implications for waste disposal in arid settings and a brief overview of important areas for future research.

2. TERMINOLOGY

The glossary at the end of this paper should help the reader understand many of the terms used in unsaturated-zone hydrology. Some of these terms are discussed in more detail below.

"Unsaturated zone" refers to the zone in which the pore space contains at least two phases, water and air. "Vadose zone" refers to the zone between land surface and the underlying aquifer. Although the terms "unsaturated zone" and "vadose zone" are generally used interchangeably, "unsaturated zone" may not be strictly accurate in some cases where perched water (which includes saturated zones) accumulates above impeding layers in an otherwise unsaturated zone. The more general term "vadose zone" may be preferred in these cases, or "variably saturated" can be used to overcome this problem.

Some classifications of arid/semiarid/humid regions have been based on mean annual precipitation (hyperarid, 0–50 mm; arid, 50–200 mm; semiarid, 200–500 mm; and humid, >500 mm [*Lloyd*, 1986]), whereas others classify regions on the basis of precipitation/evaporation ratios [*Potter*, 1992] (arid, <0.5; semiarid, 0.5–1.0; and humid, >1.0). These classifications give some idea of what is meant by "arid" and "semiarid." The term "recharge" has been generally used to describe downward water movement in the unsaturated zone; however, in thick unsaturated sections where water is moving slowly, it may be impossible to determine whether downward moving water in the upper 10–20 m will recharge the aquifer at depths ≥100 m. To avoid this problem, we use "infiltration" to refer to water movement from the surface into the subsurface and "percolation" or "drainage" to refer to penetration of water below the shallow subsurface, where most evapotranspiration occurs. "Recharge" is restricted to situations where it is likely that the water reaches the water table (shallow water table or high water flux). Although the terms "percolation" and "recharge" imply downward water movement, determining the direction of water movement is often difficult. In these situations, "water flux" is better because it implies no particular direction.

3. TECHNIQUES FOR EVALUATING WATER FLOW

Because many reviews of techniques for evaluating water flow in arid regions exist [*Edmunds et al.*, 1988; *Allison et al.*, 1994; *Phillips*, 1994], this section is not a comprehensive review of techniques. Many issues related to unsaturated flow in arid systems result from limitations of techniques used to evaluate flow; therefore a review of the limitations and assumptions associated with these techniques is important.

Techniques that are used to quantify water fluxes can be generally subdivided into physical and chemical tracer techniques. Most studies are restricted to application of one of these techniques, and although few studies apply both, use of physical and tracer methods together can provide a more comprehensive understanding of water flow. The physical approach provides an understanding of current processes, whereas chemical tracers provide information on current and long-term net water flux. Because of inherent difficulties in quantifying low water fluxes that are characteristic of many arid sites, it is important to use multiple, independent lines of data to examine unsaturated-flow processes.

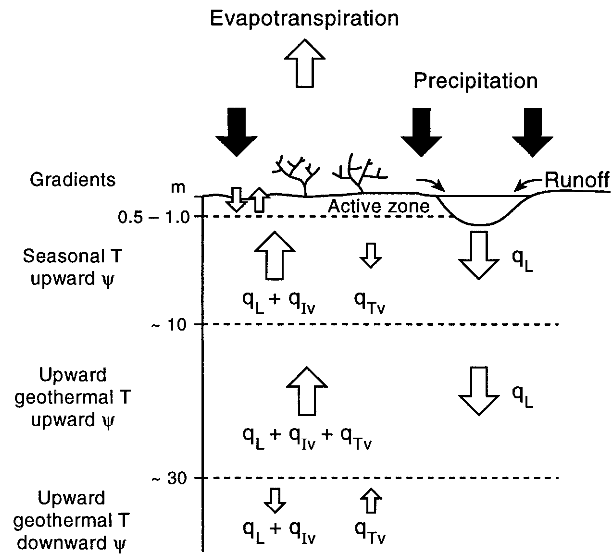

Figure 1. Schematic of unsaturated water fluxes in relation to different driving forces with depth. T is temperature, ψ is water potential, q_L is liquid water flux, q_{Iv} is isothermal vapor flux, and q_{Tv} is thermal vapor flux.

3.1. Physical Techniques

Physical techniques include water budgets to estimate water fluxes. The water balance equation can be represented by

$$D = P - R_0 - \mathrm{ET_a} - \Delta S \tag{1}$$

where D is drainage or percolation, P is precipitation (includes rain and snow), R_0 is surface runoff, $\mathrm{ET_a}$ is actual evapotranspiration, and ΔS is change in water storage (Figure 1). ET is used to describe the combined processes of evaporation (conversion of water to vapor) from the soil and transpiration from the plants. Significant improvements have been made in measuring evapotranspiration [*Malek et al.*, 1990; *Nichols*, 1994; *Albertson et al.*, 1995]; however, measurements of the different components of the water budget are generally too imprecise (±5% for P; ± 10% for $\mathrm{ET_a}$) to allow confidence in calculating the difference between numbers of nearly equal value (such as precipitation and evapotranspiration) to estimate drainage as shown by *Gee and Hillel* [1988].

Lysimeters, used to measure components of the water budget, have an artificially enclosed volume of unsaturated material [*Brutsaert*, 1982; *Allen et al.*, 1991; *Young et al.*, 1996]. Traditional lysimeters generally consist of round or square tanks that range from 1 to 5 m^2 in area and from 1 to 4 m in depth that are filled with disturbed or undisturbed soil that may be vegetated. Nonweighable lysimeters simply measure the drainage rate or amount of water percolating from the base of the lysimeter. Water storage changes can be estimated in these lysimeters by monitoring water content with a neutron probe or other device. Precipitation can be measured with a rain gauge. Most lysimeters have a rim around the surface to prevent runoff. Weighable lysimeters measure precipitation, storage changes, and drainage directly, and in this way evapotranspiration may be calculated over time spans as short as 15 min. Lysimeter measurements are considered to provide the best determination of actual evapotranspiration and are used to compare other techniques.

Lysimeter data provide valuable insights into the effects of vegetation and sediment on water movement at different sites [*Gee et al*., 1994; *Wing and Gee*, 1994]. Deep (18 m), nonweighable lysimeters at the Hanford, Washington, site measured drainage below the root zone [*Gee et al*., 1994]. To overcome the problem of limited areal extent associated with the individual lysimeters just described, large-pan lysimeters (92–322 m^2) were installed beneath engineered cover systems at the Hanford site to monitor drainage with a precision of ± 2 mm [*Tyler et al*., 1997]. Disadvantages of lysimeter studies include expense of construction, time required for maintenance, limited areal extent, boundary effects, and disturbance of the natural system. The large-pan lysimeters overcome the areal limitation, however, and when they are installed to evaluate engineered cover systems, disturbance of the natural system is not an issue.

3.1.1. Water content. Water content of sediment or rock samples can be measured readily in the laboratory by weighing samples before and after oven drying (the gravimetric method) [*Gardner*, 1986]. Because samples are destroyed during processing, this technique is generally used for collecting baseline data, for one-time routine measurements, and for calibration of other methods. It is used generally for evaluating spatial variability in water content, but not as readily for examining temporal variability. Traditionally, water content has been monitored with a neutron probe (Figure 2a) [*Gardner*, 1986], which is placed in an access tube that is installed horizontally or vertically. The neutron probe emits high-energy neutrons that collide with hydrogen nuclei and are slowed and reflected back to the probe, where they are counted. Neutron probes are calibrated against laboratory-measured water content of sediment or rock samples taken around neutron probes in the field. Calibrations are stable, and neutron probes are robust (both important for long-term monitoring). Disadvantages of neutron probes include health hazards associated with a radioactive source, time required for monitoring (generally done manually), and difficulty of monitoring the near-surface zone (top 0.15 m). Long-term (9 years) monitoring of water content was conducted in the Chihuahuan Desert, New Mexico, to evaluate spatial and temporal variability in water content [*Wierenga et al*., 1987]. Results of the monitoring show that in 8 of the 9 years, all precipitation was taken up by plant roots in the upper 1.3 m and lost by evapotranspiration back into the atmosphere.

More recently, developments in time domain reflectometry (TDR) have led to its increased use in monitoring water content (Figure 2b) [*Dalton*, 1992]. A time domain reflectometry system consists generally of a two- or three-rod probe that is connected through a transmission line to a reflectometer, such as the Tektronix 1502B (Tektronix Inc., Redmond, Oregon), at the surface. A high-frequency pulse is applied by the reflectometer to the probe or waveguide, and reflections at the beginning and end of the probe caused by impedance changes are analyzed and displayed by the reflectometer. The time required for the electromagnetic pulse to travel along the waveguide is determined by the dielectric properties of the unsaturated medium. The TDR system measures the transit time t of the pulse along the TDR probe, and the dielectric constant ε is calculated as

$$\varepsilon = (ct/2l)^2 \tag{2}$$

where c is the velocity of light in a vacuum (3×10^8 m s^{-1}) and l is the probe length. Because of large differences in the dielectric constant of water (~80), sediment or rock (~4–8), and air (~1), the dielectric constant of the unsaturated medium is controlled largely by the water content. Although *Topp et al*. [1980] developed an empirical third-order polynomial relationship between water content and dielectric constant that applies to many different sediment textures, individual calibrations can also be developed for different sediments. The average water content along the length of the TDR probe is measured. TDR probes can be installed vertically to measure average water content to a particular depth or horizontally to monitor movement of wetting fronts. A typical probe uses 0.3-m-long rods, ~5 mm in diameter, and ~20-mm spacing between rods (Campbell Scientific Inc., Logan, Utah). The advantages of TDR systems are the absence of a radioactive source, automated water content monitoring that can be operated remotely, and the ability to monitor the near-surface zone. Although TDR has not been widely implemented in arid settings, the automated measurement of water content by TDR should lead to large databases that document water content changes in arid regions.

Remote sensing has also been used to estimate water content in the unsaturated zone [*Jackson*, 1993]. This technique is based on variations in the dielectric constant with water content in unsaturated material, which is similar to that described for TDR measurements. Passive microwave remote sensing detects water content in the upper 50 mm of the unsaturated zone at a spatial resolution of ~200 m [*Jackson et al*., 1993]. The shallowness of the zone being evaluated and the low spatial resolution make this technique unsuitable for evaluation of unsaturated-zone water fluxes at small scales; it is generally more applicable in basin-scale studies and climate modeling.

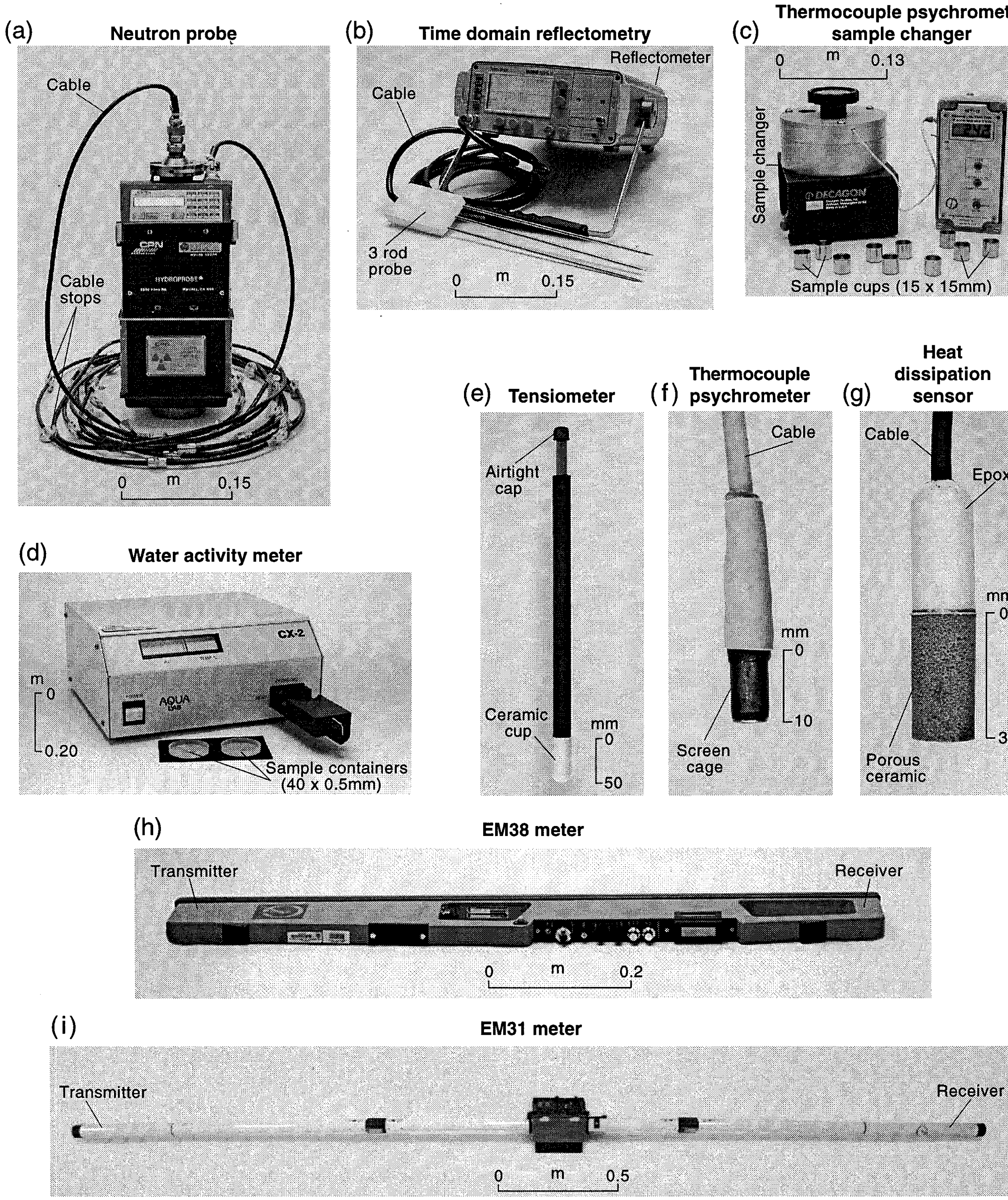

Figure 2. Instrumentation used for monitoring various parameters in the unsaturated zone: (a) neutron probe (model CPN 503DR), (b) time domain reflectometry system (reflectometer and three-rod probe), (c) thermocouple psychrometer sample changer, (d) water activity meter, (e) tensiometer, (f) thermocouple psychrometer, (g) heat dissipation sensor, (h) EM38 meter, and (i) EM31 meter.

Spatial variability in water content cannot be used to evaluate water flux in heterogeneous systems because water content varies with sediment type: clays, for example, retain more water than do sands. In contrast, temporal variations in water content can be used to evaluate the movement of water pulses through the unsaturated zone, particularly in areas of moderate to high water flux; however, in areas of low water flux, typical standard

TABLE 1. Summary of Instruments Used to Measure Various Hydraulic Parameters in Arid Systems

Parameter	*Instrument*	*Range*	*Accuracy*	*Notes*
Water content	neutron probe	0 to 100% saturation	±1%	robust, radioactive source
	TDR	0 to 100% saturation	±1%	robust, nonradioactive, automated
Matric potential	HDS	−0.01 to −1.4 MPa		not robust, automated
	tensiometer	0 to −0.08 MPa		automated
Water potential	TCP	−0.2 to −8.0 MPa	±0.2 MPa	not robust, automated
	Filter paper	−0.2 to −90 MPa		laboratory measurement
	SC10A sample changer	−0.2 to −8.0 MPa (Peltier)	±0.2 MPa	laboratory measurement affected by temperature
		−0.2 to −300 MPa (Spanner)	±0.2 MPa	gradients; time consuming
	water activity meter	0 to −312 MPa	±0.003 activity units	rapid laboratory measurement
Hydraulic conductivity	centrifuge method	$\geq 10^{-11}$ m s^{-1}	~±10%	expensive

Abbreviations are TDR, time domain reflectometry; HDS, heat dissipation sensor; and TCP, thermocouple psychrometer.

errors (~±1% for calibration curves for instruments (Table 1)) associated with water content measurements at one location over time may be too high to detect low water fluxes. Water content cannot be used to estimate water flux under steady flow conditions because water content does not vary.

3.1.2. Potential energy. In contrast to water content, which cannot be used to evaluate the direction of water movement because water content is discontinuous across the interface between different sediment textures, potential energy can be used to assess the direction of the driving force for water movement. Water flows from regions of high potential to regions of low potential. Potential energy in the unsaturated zone includes capillary, adsorptive, gravitational, solute or osmotic, and pneumatic components (Table 2). Capillary and adsorptive components combine to form the matric potential, which is the component of potential energy associated with the matrix of the unsaturated zone. The term "matrix" describes the particles and pore space that make up the unsaturated medium; "matric" is its adjectival form (Webster's Third International Dictionary). "Gravitational potential" represents the elevation of the measurement point above a reference level, such as the water table. Solute or osmotic potential results from the reduction in energy associated with addition of solutes to pore water. Matric and osmotic components are combined to form water potential. Because osmotic potential is generally neglected except in cases where high solute concentrations exist, "water potential" and "matric potential" are often used interchangeably. Pneumatic potential results from changes in air pressure in the unsaturated zone. Potential energy is generally expressed as energy per unit volume (pressure equivalent in megapascals) or energy per unit weight (head equivalent in meters).

The pore space in unsaturated media is partially filled with water, and pressures are negative. Matric potentials and water potentials are negative, whereas suction or tension, the negative of the matric potential, is positive (Table 2). The general term "pressure potential" is used in this paper, along with more appropriate, specific terms for clarity. Pressures close to 0 correspond to near-saturated conditions, and low negative pressures correspond to dry conditions. Water flows from regions of high potential, where pressures are less negative, to

TABLE 2. Various Types of Potential Energy Important for Understanding Unsaturated Flow

Potential Energy Type	*Description*
Gravitational potential	elevation above reference level (e.g., water table)
Matric potential	capillary and adsorptive forces associated with the soil matrix
Suction or tension	negative matric potential
Osmotic (solute) potential	variations in potential energy associated with solute concentration
Water potential	matric + osmotic potential
Pneumatic potential	associated with variations in air pressure
Hydraulic head	matric + gravitational potential head

Water potential approximates matric potential when osmotic potential is negligible. Tensiometers generally measure matric potential because air pressure is usually atmospheric. Heat dissipation sensors measure matric potential. Thermocouple psychrometers measure water potential. Potential energy is generally expressed as energy per unit weight of water, which is equivalent to head (meters) or energy per unit volume of water, which is equivalent to pressure (megapascals).

regions of low potential, where pressures are more negative.

Tensiometers (Soil Measurement Systems, Tucson, Arizona; Soil Moisture Equipment Corporation, Santa Barbara, California) can be used to monitor high (≥ -0.08 MPa) pressure potentials (matric and pneumatic) generally found in humid sites; however, pressure potentials in arid sites have a wide range (0 to < -200 MPa), and thus tensiometers can only be used where the vadose zone is relatively moist (Figure 2e; Table 1). Tensiometers consist of a ceramic cup connected to an airtight PVC tube that is filled with water (Figure 2e) [*Cassel and Klute*, 1986]. Water in the tensiometer equilibrates with the surrounding unsaturated medium, and a vacuum is developed that is measured by a pressure transducer.

Heat dissipation sensors (Campbell Scientific Inc., Logan, Utah) (Figure 2g), also called matric potential sensors, measure matric potential over a range (-0.01 to -1.4 MPa) greater than that of tensiometers [*Campbell and Gee*, 1986; *Phene et al.*, 1992]. Heat dissipation sensors consist of a ceramic block, a heater, and a temperature transducer. Heat dissipation sensors (1) measure the matric potential of the unsaturated medium by equilibrating a standard matrix, such as porous ceramic, with the surrounding sediments and (2) determine the water content of the sensor by measuring the rate of heat dissipation, which is a function of water content of the ceramic block. The higher the water content of the soil and the less negative the matric potential, the more rapidly the heat dissipates, and the lower the recorded voltage. The temperature change is measured with a data logger before and after application of a 30 s heat pulse. Temperature measurements are related to matric potentials through calibration curves between temperature or voltage and matric potential measured in the laboratory. Because matric potential is continuous across material types, the matric potential of the heat dissipation sensor is the same as that of the surrounding unsaturated medium [*Thamir and McBride*, 1985].

Thermocouple psychrometers (J.R.D. Merrill Specialty Co., Logan, Utah; Wescor, Logan, Utah) are required to measure much more negative water (matric + osmotic) potentials associated with typically dry sediments in arid systems. Thermocouple psychrometers measure the relative humidity of the vapor phase in the unsaturated zone, which is related to the water (pressure) potential in the liquid phase, according to the Kelvin equation

$$\psi = \frac{RT}{V_w} \ln \frac{P}{P_0} \tag{3}$$

where R is the ideal gas constant (8.314 J mol^{-1} °K), T is the Kelvin temperature, V_w is the molar volume of water (1.8×10^{-5} m^3 mol^{-1}), and P/P_0 is the relative humidity expressed as a fraction (P is the vapor pressure of the air in equilibrium with the sample, and P_0 is the saturation vapor pressure) [*Rawlins and Campbell*, 1986]. There are two basic types of thermocouple psychrometers: (1) Peltier or Spanner psychrometers (Figure 2f) and (2) Richards psychrometers (Figure 2c). Peltier psychrometers consist of a small thermocouple junction in a sample chamber such as the screen cage in Figure 2f that is cooled by the Peltier effect to condense water on it. The Richards psychrometer mechanically adds a drop of water to the thermocouple junction that is within the sample chamber (Figure 2c) and is restricted to laboratory measurements. Both systems measure temperature depression of the wet, or measuring, junction relative to a dry, or reference, thermocouple junction in the chamber. Temperature depression varies with the rate of evaporation, which is greater at lower relative humidity. A primary source of error results from temperature gradients between the reference junction and pore water in the unsaturated zone. A temperature gradient of 1°C at 20°C results in an error in measured water potentials of 13 MPa [*Rawlins and Campbell*, 1986]. Thermocouple psychrometers are calibrated with salt solutions of known osmotic potential.

In situ thermocouple psychrometers (Figure 2f) are used to monitor water potential between -0.2 and -8.0 MPa. Water potentials have been monitored in various arid settings to a maximum depth of 387 m to evaluate the direction of water movement and to estimate water fluxes [*Montazer et al.*, 1985; *Fischer*, 1992; *Scanlon*, 1994]. Significant improvements have been made in thermocouple psychrometry for monitoring water potentials in the field in recent years as a result of advances in data acquisition systems and newly developed thermocouple psychrometers for installation in deep boreholes [*Kume and Rousseau*, 1994].

One problem inherent in monitoring pressure potentials in arid systems is that the installation process may significantly affect the natural system, causing the monitoring data to be an artifact of the installation process rather than a reflection of the natural system. Although thermocouple psychrometers are generally installed in dry materials, because equilibration of the backfill sediments may take a long time, determining the true potential of the sediments may be difficult. Numerical simulations conducted to examine the effect of borehole backfill on monitored water potentials in a fractured tuff site show that backfill material could greatly disturb the natural system [*Montazer*, 1987]. Heat dissipation sensors are generally installed in wet silica flour because they require good contact with the surrounding sediment [*Montazer et al.*, 1985]; however, measured discrepancies between closely spaced thermocouple psychrometers and heat dissipation sensors suggest that the wetted sediments may not equilibrate for a long time. Because the calibration is unstable and because the instruments are not robust and have a high failure rate, thermocouple psychrometers may be unsuitable for long-term (≥ 10 years) monitoring unless they are retrievable. Installation of retrievable thermocouple psychrometers in caissons and in boreholes [*Fischer*, 1992; *Prudic*, 1994] has allowed recalibration of these instruments.

Because of the expense and difficulties of installing thermocouple psychrometers in the field, we generally obtain information on spatial variability of water (pressure) potential on the basis of laboratory measurements on disturbed samples by using a thermocouple psychrometer with a sample changer (Figure 2c) or a water activity meter (Decagon Devices, Pullman, Washington) (Figure 2d). The sample changer uses a Richards thermocouple psychrometer to measure a wide range in water (pressure) potentials (−0.2 to −300 MPa [*Rawlins and Campbell*, 1986]). Laboratory measurements of water potential made by thermocouple psychrometers are time-consuming and sensitive to the effects of temperature gradients [*Rawlins and Campbell*, 1986].

A water activity meter (Figure 2d) can also be used to measure water (pressure) potential in the laboratory. Water activity is synonymous with relative humidity. Water potential measurements made by a water activity meter are neither as time-consuming nor as sensitive to the effect of temperature gradients as are measurements made by thermocouple psychrometers [*Gee et al*., 1992]. The measurement of water activity of a sediment or rock sample takes only a few minutes, ranging from 0.100 to 1.000 (−312 to 0 MPa water potential) with uniform resolution of ±0.003 water activity units throughout the range [*Gee et al*., 1992]. The water activity meter uses a chilled mirror to measure the dew point of water vapor above a small sample of sediment or rock (40 mm in diameter by 5 mm thick). A Peltier cooling device controlled by a data logger is used to cool the mirror until dew forms and then to heat the mirror to eliminate the dew. Temperature of the sediment or rock sample is measured with an infrared thermometer. Vapor pressure of air is equal to the saturation vapor pressure at the dew point temperature, by definition of the dew point. Saturation vapor pressure is approximated by

$$P_0(T) = a \exp\left(\frac{bT_s}{T_s + c}\right) \tag{4}$$

where a, b, and c are constants and T_s is the surface temperature [*Buck*, 1981].

$$\begin{aligned} A_w = \frac{P}{P_0(T_s)} &= a \exp\left(\frac{bT_d}{T_d + c}\right)\left[a \exp\left(\frac{bT_s}{T_s + c}\right)\right]^{-1} \\ &= \exp\left(\frac{bT_d}{T_d + c} - \frac{bT_s}{T_s + c}\right) \\ &= \exp\left(\frac{bc(T_d - T_s)}{(T_d + c)(T_s + c)}\right) \end{aligned} \tag{5}$$

where T_d is the dew point temperature in degrees celsius. A microprocessor-controlled algorithm is used to convert the air dew point temperature and the sample temperature to water activity or relative humidity readings. The Kelvin equation (equation (3)) is then used to estimate the water potential. Temperature control is unimportant because change in water activity with temperature is generally $<0.003°\text{C}^{-1}$. Because the chilled mirror dew point technique is a primary measurement method of relative humidity, no calibration is required.

The filter paper method, also used to measure matric or water potentials on sediment or rock samples in the laboratory ranging from −0.2 to −90 MPa, does not require expensive instrumentation [*Greacen et al*., 1987; *American Society for Testing and Materials*, 1994]. This method assumes that porous media in liquid or vapor contact with the filter paper will exchange water until the matric or water potentials of both are the same. The filter paper can be placed in direct contact with the sample to measure the matric potential, or it can be separated from the sample by a vapor gap to measure water potential (matric and osmotic potential). Although the time required for equilibration varies with the potential of the medium, equilibrium is generally reached within 7 days. Whatman no. 42 filter papers are generally used, and the increase in mass of the filter paper is measured and related to matric or water potential through a previously determined calibration curve. *Greacen et al*. [1987] listed calibration equations for different ranges in water potential. The greatest source of error in all laboratory measurements of pressure (water or matric) potentials is the possibility of samples drying during collection, particularly in coarse-textured material.

3.1.3. Hydraulic conductivity. Information on hydraulic conductivity is required for estimating water flux using Darcy's law under steady flow conditions or using Richards' equation under transient flow conditions. Darcy's law is empirical and was originally developed for the saturated zone. Darcy's law shows that water flux under steady flow is proportional to the hydraulic head gradient, the proportionality constant being the hydraulic conductivity. Hydraulic head is the sum of the matric (pressure) potential head and the gravitational potential head. In the saturated zone, hydraulic conductivity is constant at a point in space. Darcy's law was modified by *Buckingham* [1907] for the unsaturated zone by allowing the hydraulic conductivity K to vary with water content θ:

$$q_l = -K(\theta)\frac{\partial H}{\partial z} = -K(\theta)\left(\frac{\partial h(\theta)}{\partial z} + 1\right) \tag{6}$$

where q_l is the liquid water flux, H is the hydraulic head, and h is the matric potential head, which is a function of the water content. Richards' equation is required to predict water content or matric potential in the unsaturated zone during transient flow and combines the conservation of mass with Darcy's equation (conservation of momentum):

$$\frac{\partial\theta}{\partial t} = -\frac{\partial q_l}{\partial z} = \frac{\partial}{\partial z}\left[K(\theta)\left(\frac{\partial h(\theta)}{\partial z} + 1\right)\right] \tag{7}$$

Although unsaturated hydraulic conductivity is the least well known flow parameter, it has a great effect on estimated water fluxes because hydraulic conductivity may vary over several orders of magnitude in the range of water contents found in arid regions. Unsaturated hydraulic conductivity can be estimated from water retention and saturated hydraulic conductivity data by assuming that the unsaturated medium behaves like a bundle of capillary tubes [*Mualem*, 1976; *van Genuchten*, 1980]; however, in many arid regions, water may be adsorbed as films, and estimates of hydraulic conductivity based on capillary flow may not apply.

There are numerous field and laboratory methods for determining the unsaturated hydraulic conductivity as a function of water content. These methods are either steady state or transient and are described in detail by *Klute* [1986]. Recent developments in ultracentrifuge technology allow measurement of unsaturated hydraulic conductivity at fairly low water contents. *Nimmo et al.* [1987, 1992] and *Conca and Wright* [1992] developed steady state centrifuge methods to measure unsaturated hydraulic conductivity. Large forces (≤2000 g per unit mass) applied to the unsaturated sample result in removal of water from the sample. The magnitude of the force is controlled by the radius and speed of rotation of the centrifuge [*Kutilek and Nielsen*, 1994]. The various centrifuge methods apply water at a constant rate to the inner side of a small sediment sample or rock core either through precision pumps or through a water reservoir and porous ceramic plate. The sample generally reaches a steady state water content in a fairly short time. The steady state water flux can be described by a modified Darcy equation:

$$q_1 = -\frac{K(\theta)}{g}\left(\frac{dh(\theta)}{dr} - \omega^2 r\right) \qquad (8)$$

where K is the unsaturated hydraulic conductivity, g is the gravitational acceleration, r is the radius of the sample, and $\omega^2 r$ is the centripetal force per unit mass. Assuming a negligible or unit gradient ($dh/dr = 1$), the unsaturated hydraulic conductivity is calculated by dividing the measured flux q_1 by $\omega^2 r g^{-1}$. The sample is removed from the centrifuge, and the water content and/or matric potential is measured. The experiment is rerun at different flow rates to calculate the unsaturated hydraulic conductivity at different water contents or matric potentials.

3.1.4. Noninvasive techniques for estimating water content and movement. Because of the difficulties and expense of installing dedicated equipment, particularly in contaminated sites, noninvasive techniques for evaluating unsaturated water movement are highly desirable. In disposal sites, equipment installation should be minimized to maintain site integrity and to avoid creating preferred pathways for contaminants.

Electromagnetic induction (EMI) has been used to evaluate spatial variability in unsaturated flow over large regions [*Cook et al.*, 1992; *Cook and Kilty*, 1992]. EMI is a noninvasive technique that measures apparent electrical conductivity, which can be used to evaluate unsaturated flow. The theoretical basis for electromagnetic induction measurements is described by *McNeill* [1992]. The instruments (e.g., EM38 meter (Figure 2h) or EM31 meter (Figure 2i), Geonics Inc., Mississauga, Ontario) generally consist of a transmitter coil placed on the ground that is energized by an alternating current at an audio frequency. This current generates a primary magnetic field, which in turn induces small currents that generate their own secondary magnetic field. The receiver coil responds to both the primary and secondary magnetic field components. Under low values of induction number, the secondary magnetic field is a linear function of apparent electrical conductivity. The instrument can be operated with both transmitter and receiver coils lying horizontally (vertical dipole mode) or vertically (horizontal dipole mode) on the ground.

Ground-based EMI surveys can be conducted with a variety of instruments that range in exploration depth from 0.75 to 40 m (Figure 2) [*McNeill*, 1992]. Apparent electrical conductivity (EC_a) in the subsurface is related to water content, salt content, texture, structure, and mineralogy:

$$EC_a = EC_w \theta \tau + EC_s \qquad (9)$$

where EC_w is pore-water conductivity, θ is volumetric water content, τ is tortuosity, and EC_s is surface conductance of the sediment [*Rhoades et al.*, 1976]. Higher recharge generally occurs in more coarsely textured soils (lower EC_a) and results in higher relative water content (higher EC_a) and lower chloride content (lower EC_a) [*Cook et al.*, 1992]. Because of competing effects of texture, chloride, and water content on EC_a, EMI will work well only in recharge estimation where any one of these factors dominates or where two factors operate synergistically on EC_a. In an Australian study, because the correlation between recharge and EC_a was controlled by soil texture, the EMI survey mapped primarily soil texture at the site [*Cook et al.*, 1992]. Comparison of ground measurements of EC_a with recharge estimated according to unsaturated-zone chloride data at 20 sites resulted in a coefficient of determination (R^2) of 0.5. These data suggest that although EMI cannot estimate recharge directly, it may be useful in reconnaissance and interpolation between borehole measurements.

An electromagnetic meter (Geonics model EM31 (Figure 2i)) has also been used to monitor temporal variations in water content along an ~2-km transect [*Sheets and Hendrickx*, 1995]. The researchers found a linear relationship between apparent conductivity measured using the EM31 meter and water content in the upper 1.5 m of soil logged in 65 neutron probe access tubes along the transect. This technique shows promise for monitoring water content in disposal facilities, once a calibration equation has been developed.

TABLE 3. Summary of Environmental Tracers Commonly Used in Arid Regions and Their Attributes

Tracer	*Type*	*Liquid/Vapor Phase*	*Dating Period, years*	*Notes*
Chloride		liquid	≤1000s	qualitative
^{36}Cl	bomb pulse	liquid	0–40	used in evaluating water fluxes and preferential flow
	cosmogenic variation	liquid	≤70,000	small signal ≤2 × background; advection-dominated systems
	radioactive decay	liquid	50,000–1,000,000	used at Yucca Mountain, Nevada
^{3}H	bomb pulse	liquid + vapor	0–40	used in evaluating water fluxes and preferential flow

3.2. Tracer Techniques for Estimating Water Movement

It is difficult to estimate rates of water movement in unsaturated media because the rates are generally low. Physical methods that depend on Darcy's or Richards' equations are restricted by uncertainties in estimated unsaturated hydraulic conductivities. Chemical tracers can provide information on current water fluxes and long-term net water fluxes for up to thousands of years. In humid sites, applied tracers (such as bromide) are used for evaluating solute transport. Organic dyes (such as FD&C (food, drug, and cosmetics) blue dye and Rhodamine dye) have also been used in delineating preferred pathways in humid regions [*Steenhuis et al.*, 1990]. Use of applied tracers has generally been limited in arid regions to irrigated areas [*Wierenga et al.*, 1991] or localized zones of high water fluxes [*Scanlon*, 1992b]. The low water fluxes typical of many arid settings limit the penetration depth of applied tracers. In some arid settings, contaminants in the unsaturated zone can be considered long-term applied tracers. Bromide that originated in a factory that had been operating for 18 years was used to evaluate water flow and solute transport at a site in the Negev Desert, Israel [*Nativ et al.*, 1995].

A wide variety of environmental tracers exists that span different time scales (Table 3). These tracers, including ^{36}Cl and ^{3}H, are produced naturally in the Earth's atmosphere and have existed in the natural environment for millions of years. The concentration of these tracers was greatly increased by nuclear testing in the mid-1950s to early 1960s, however (Figure 3). Some tracers exist in both liquid and vapor phases (tritiated water), whereas others exist only in the liquid phase in the subsurface (Cl and ^{36}Cl). We will review some of the most widely used environmental tracers and examine the assumptions associated with these tracers and how accurately they represent the flow system.

3.2.1. Meteoric chloride. The chloride mass balance approach uses chloride concentrations in pore water to estimate liquid water fluxes for up to thousands of years at many arid sites [*Allison and Hughes*, 1983]. Chloride from precipitation, dry fallout, or irrigation may concentrate in the root zone as a result of evapotranspiration [*Gardner*, 1967]. Chloride transport through the unsaturated zone is described by

$$q_{\mathrm{Cl}} = q_{\mathrm{l}} c_{\mathrm{Cl}} - D_{\mathrm{h}} \frac{\partial c_{\mathrm{Cl}}}{\partial z} \tag{10}$$

where q_{l} is the volumetric liquid water flux below the root zone (L T^{-1}), q_{Cl} is the chloride deposition flux at the surface (M L^{-2} T^{-1}), c_{Cl} is the pore water chloride concentration (M L^{-3}), and D_{h} is the hydrodynamic dispersion coefficient (L^2 T^{-1}), a function of θ (volumetric water content) and v (average pore water velocity). The first term on the right represents the chloride flux that results from advection, and the second term represents the flux from hydrodynamic dispersion. The mechanical dispersion coefficient D_{m} and the effective molecular diffusion coefficient D_{e} compose the hydrodynamic dispersion coefficient. Mechanical dispersion is the mixing that occurs as a result of variations in pore water velocity due to (1) the parabolic velocity distribution within a pore, (2) different pore sizes, and (3) the effects of tortuosity or branching of pore channels. Molecular diffusion results from the thermal or kinetic energy of particles. Mechanical dispersion is assumed to be negligible because flow velocities are generally <7 m yr^{-1}, which *Olsen and Kemper* [1968] specified as the water velocity below which mechanical dispersion can be ignored. The effective molecular diffusion coefficient differs from the diffusion coefficient in pure water because of the reduced cross-sectional area in unsaturated media (represented by the water content)

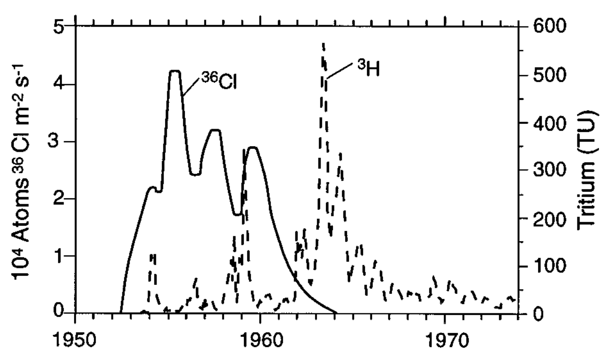

Figure 3. Temporal variations in predicted bomb ^{36}Cl fallout between 30°N and 50°N latitude [*Bentley et al.*, 1986] and in ^{3}H fallout of precipitation in the northern hemisphere [*IAEA*, 1983], decay corrected to 1989.

and the increased path length for the water (tortuosity). At low water fluxes the diffusive flux may be dominant. In many arid systems the hydrodynamic dispersion coefficient can be assumed to be negligible [*Allison and Hughes*, 1978], and equation (10) is simplified to

$$q_l = q_{Cl}/c_{Cl} \tag{11}$$

The age of the chloride and, by implication, that of the water can be calculated by dividing the integrated Cl content from the surface to the depth of interest by the annual chloride deposition flux. Chloride concentration in pore water is inversely proportional to water flux: low chloride concentrations indicate high water flux, and high chloride concentrations indicate low water flux (Figure 4).

Chloride deposition flux at a site can be estimated by (1) measuring chloride concentrations in precipitation and dry fallout or (2) dividing the natural ^{36}Cl fallout at a site, which varies according to latitude (as predicted by *Andrews and Fontes* [1991]), by the prebomb $^{36}Cl/Cl$ ratio (i.e., ratios before the first atmospheric nuclear explosion). An independent estimate of chloride deposition was also calculated for chloride profiles at the Hanford site, Washington [*Murphy et al.*, 1996]. Late Pleistocene floods, resulting from breaching of glacial dams, reset the chloride mass balance clock at the beginning of the Holocene. Estimates of chloride deposition that were calculated by dividing the chloride mass by the time since flooding when all chloride was flushed out of the sediments (15,000 years) agreed with estimates based on prebomb $^{36}Cl/Cl$ ratios. Because chloride mass balance equations are linear, uncertainties in the chloride deposition flux result in corresponding uncertainties in estimated water fluxes. If chloride concentration in precipitation is controlled (to first order) by distance from the ocean, its concentration should not vary significantly with time. Higher precipitation during Pleistocene times would result in correspondingly higher chloride deposition. Chloride deposition from dry fallout of dust and salts is of the same magnitude as that from precipitation in Nevada [*Dettenger*, 1989]. The contribution of dry fallout from saline lakes can be examined by measuring prebomb $^{36}Cl/Cl$ ratios because saline lakes have signatures markedly different from those of modern precipitation [*Phillips et al.*, 1995]. The prebomb $^{36}Cl/Cl$ ratios refer to $^{36}Cl/Cl$ ratios at depth that reflect fallout that occurred before the bomb pulse. At many sites, prebomb $^{36}Cl/Cl$ ratios are similar (500×10^{-15} [*Scanlon*, 1992a; *Fabryka-Martin et al.*, 1993]), which suggests that the contribution of ^{36}Cl from saline lakes is negligible at these sites. In addition to rain and dry fallout, other sources of chloride include rocks at Yucca Mountain [*Fabryka-Martin et al.*, 1993] and runon or runoff that should be quantified.

The chloride mass balance approach assumes piston-like flow, or uniform downward movement of water that displaces the initial water in the profile. The assumption

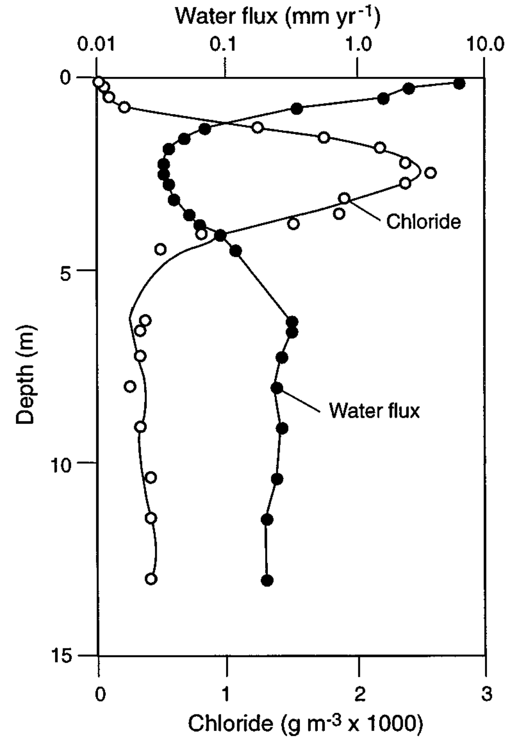

Figure 4. Typical example of inverse relationship between pore water chloride concentrations and estimated water fluxes. Adapted from *Scanlon* [1991, Figure 2] with kind permission from Elsevier Science–NL, Amsterdam, Netherlands.

of piston-like flow has been questioned at many sites. Because chloride input to the system is continuous, chloride profiles are generally insensitive to preferential flow, or nonuniform downward movement of water, in which some water moves rapidly along preferred pathways such as roots or fractures. Piston-like and preferential flow are discussed in more detail in section 5. Evidence of preferential flow is generally provided by the distribution of bomb pulse tracers in the vadose zone such as bomb pulse ^{36}Cl and ^{3}H. Although tritium data at a fractured-chalk site in the Negev Desert indicate preferential flow, chloride profiles at this site are smooth, as would be expected at a site without preferential flow [*Nativ et al.*, 1995].

Bulge-shaped chloride profiles at many sites in nonfractured sediments could result from preferential flow [*Nativ et al.*, 1995], diffusion to a shallow water table [*Cook et al.*, 1989], or transient flow [*Scanlon*, 1991; *Phillips*, 1994] (Figure 5). Chloride profiles at many of these sites look similar, and interpretation of the bulge shape generally relies on additional information. Evidence of preferential flow in the Negev site was provided by deep penetration of ^{3}H [*Nativ et al.*, 1995]. The shape of some profiles in Australia are attributed to diffusion to a shallow water table because of the differences in chloride concentration between unsaturated and saturated zones (Figure 5) [*Cook et al.*, 1989]. Bulge-shaped chloride profiles in the southwestern United States,

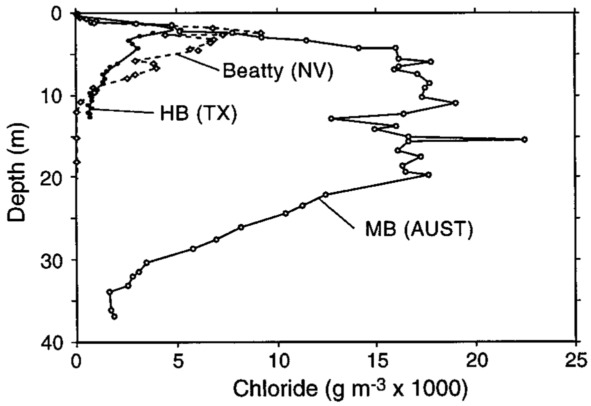

Figure 5. Bulge-shaped chloride profiles from vegetated dunes in the Murray Basin, South Australia (MB, profile BVDO1 [*Cook et al.*, 1989]), and from various southwestern U.S. settings (Hueco Bolson (HB), Texas, [*Scanlon*, 1991]; Beatty, Nevada [*Prudic*, 1994]). HB and MB plots reproduced from *Scanlon* [1991, Figure 3] and *Cook et al.* [1989] with kind permission from Elsevier Science–NL, Amsterdam, Netherlands.

where the water table is generally much deeper (≥100 m), are attributed to higher water fluxes during the Pleistocene, when the climate was cooler and wetter [*Scanlon*, 1992a; *Phillips*, 1994; *Tyler et al.*, 1996]. Additional evidence on the effect of paleoclimate on water movement is provided by stable isotopic data [*Tyler et al.*, 1996]. In areas where the chloride concentration below the chloride peak is very low, such as at Beatty, Nevada [*Prudic*, 1994], preferential flow cannot be used to explain the reduction in chloride because preferential flow refers to enhanced water movement along localized preferred pathways, which does not include complete leaching (Figure 5). Because chloride profiles represent net liquid water flux over long time periods, the chloride at depth at these sites is a relic of past climate conditions and does not represent current conditions. In Australia, on a much smaller timescale (~100 years), transient flow conditions resulted when native mallee vegetation, characterized by deep-rooted (~20 m) eucalyptus trees, was replaced by crops and pasture [*Cook et al.*, 1994].

The chloride mass balance method provides an estimate of liquid water flux, which is important in evaluating the movement of nonvolatile solutes. Because liquid water flux may move downward and vapor flux and net water flux may move upward, estimates of liquid flux based on chloride data alone may provide inaccurate estimates of net water flux.

3.2.2. Chlorine 36. Chlorine 36 (half-life of 301,000 years) is produced in the atmosphere by cosmic ray spallation of ^{36}Ar and neutron activation of ^{35}Cl [*Bentley et al.*, 1986]. Chlorine 36 can provide estimates of liquid water residence time (1) over the past ~40 years by means of bomb pulse $^{36}Cl/Cl$ ratios, (2) over the past 70–80 kyr by means of variations in cosmogenic production of ^{36}Cl, and (3) from 50 to 1000 kyr by means of radioactive decay of ^{36}Cl (Table 3).

Nuclear weapons tests conducted between 1952 and 1958 resulted in ^{36}Cl concentrations in precipitation as much as 1000 times greater than natural fallout levels [*Bentley et al.*, 1986] (Figure 3). In nonfractured sediments, water fluxes have been estimated from the ^{36}Cl center of mass [*Cook et al.*, 1994]. The amount of water in the profile above the center of mass of ^{36}Cl is equal to the flux during the time period since the center of mass of the fallout occurred. Annual water flux is generally calculated by dividing this total flux by time in years. In many areas where bomb pulse ^{36}Cl has been used to estimate water flux, the center of mass of the bomb pulse is still in the root zone [*Gifford*, 1987; *Norris et al.*, 1987; *Phillips et al.*, 1988; *Scanlon*, 1992a] (Figure 6). Occurrence of the bomb pulse in the root zone indicates that water fluxes at these sites are extremely low, which is important for waste disposal. Because much of this water in the root zone is later removed by evapotranspiration, water fluxes estimated from tracers within the root zone overestimate water fluxes below the root zone by up to several orders of magnitude [*Tyler and Walker*, 1994]. High $^{36}Cl/Cl$ ratios have been found to depths of 440 m at Yucca Mountain, Nevada [*Liu et al.*, 1995], suggesting preferential flow along fractures. Variations in cosmogenic production of ^{36}Cl during the past 60–70 kyr could complicate the use of bomb pulse $^{36}Cl/Cl$ ratios. Some of the measured $^{36}Cl/Cl$ ratios considered to be bomb pulse, particularly at Yucca Mountain, fall within the range estimated as a result of variations in cosmogenic production of ^{36}Cl (J. Fabryka-Martin, personal communication, 1995) and may not be bomb related. Because the ratio of ^{36}Cl to chloride rather than the ^{36}Cl concentration is measured, high chloride concentrations in pore water could reduce the effectiveness of $^{36}Cl/Cl$ ratios to estimate preferential flow.

Variations in cosmogenic production of ^{36}Cl can also be used to date water during the past 70–80 kyr [*Phillips et al.*, 1991; *Plummer and Phillips*, 1995]. Production rates of meteoric ^{36}Cl vary inversely with the strength of the magnetic field and increased by as much as a factor of 2 during periods of reduced magnetic field strength [*Plummer and Phillips*, 1995]. Comparison of reconstructed ^{36}Cl production with variations in ^{36}Cl in pore water has been used to estimate ages of water at the Nevada Test Site [*Tyler et al.*, 1996]. Because variations in cosmogenic production increase the background ratio by only as much as a factor of 2, such variations may not be readily preserved in the unsaturated zone because of diffusion and dispersion.

Radioactive decay of ^{36}Cl has also been used to date very old pore water in the unsaturated zone at Yucca Mountain [*Fabryka-Martin et al.*, 1993]. Use of radioactive decay of ^{36}Cl is complicated at this site because contributions of "dead" Cl (having no ^{36}Cl) from rock away from the main flow regime result in greater apparent ages.

3.2.3. Tritium. Tritium (^{3}H; half-life of 12.4 years), produced by cosmic ray neutrons interacting with

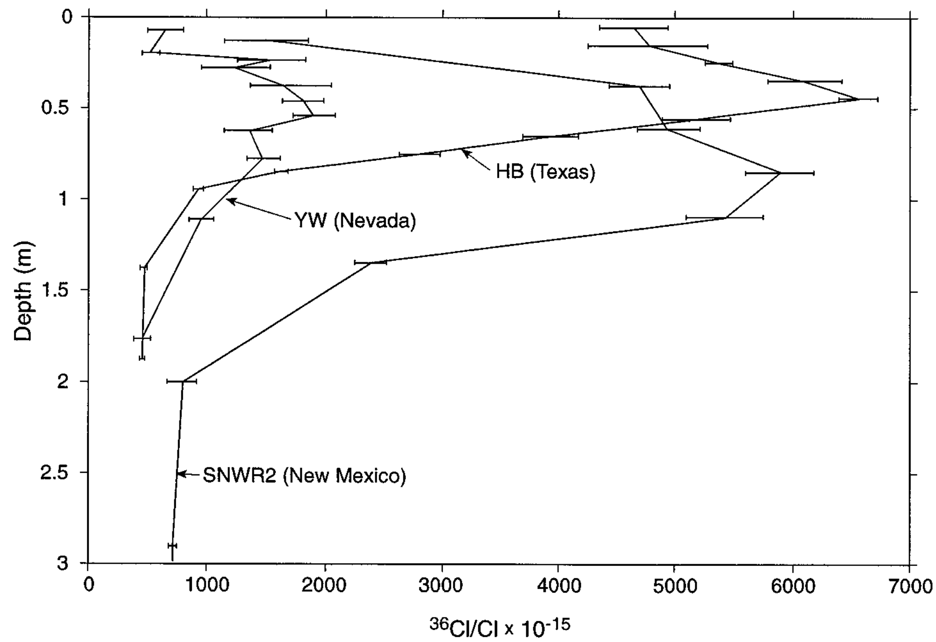

Figure 6. Profile of $^{36}Cl/Cl$ ratios from the Chihuahuan Desert (Hueco Bolson (HB) [*Scanlon*, 1992a]. Yucca Wash, Nevada (YW [*Norris et al.*, 1987]); and Sonoran Desert, New Mexico (SNWR2 [*Phillips et al.*, 1988]). Bars represent 1 standard deviation in the $^{36}Cl/Cl$ ratios. YW plot reproduced from *Norris et al.* [1987, Figure 1] with kind permission from Elsevier Science–NL, Amsterdam, Netherlands.

nitrogen in the upper atmosphere, typically results in 5–10 tritium units (TU) in precipitation. Tritium concentrations increased from 10 to ≥2000 TU during atmospheric nuclear testing [*International Atomic Energy Agency* (*IAEA*), 1983] that began in 1952 and peaked in 1963–1964 (Figure 3). Because tritiated water exists in both liquid and vapor phases, tritium is a tracer for liquid and vapor water movement. The distribution of bomb-pulse tritium in the vadose zone can be used to estimate water fluxes and to evaluate preferential flow, a procedure similar to that described for ^{36}Cl.

In tritium analysis, pore water can be extracted directly from cores by means of toluene distillation or cryodistillation. Alternatively, gas samples can be extracted from boreholes and water condensed from the gas for tritium analysis. Large gas volumes are required to detect the trace amounts of tritium found at some sites that lead to uncertainties in the volume and depth interval of the unsaturated section that is sampled. Contamination in gas sampling procedures may occur because of the potential for air flow along well casing and leaking gas lines. Problems in interpreting very low tritium levels in Ward Valley, California, a proposed low-level radioactive waste disposal facility, are thought to result from poor sampling procedures and from the absence of procedural blanks for evaluating possible contamination [*NRC*, 1995]. General problems with analysis of low tritium levels in the unsaturated zone, particularly those close to the detection limit, may reflect our lack of experience with environmental tritium sampling and our inability to collect reliable samples.

To analyze water samples for tritium, various techniques have been used that depend on the amount of water available for analysis and accuracy required. Direct liquid scintillation generally requires ~20 mL of water, and the detection limit is ~6 TU (C. Eastoe, personal communication, 1995). The detection limit is greatly reduced when electrolytic enrichment is used; however, the amount of water required is greater. A minimum sample size of 275 mL, an electrolytic enrichment factor of ~80, and a counting time of 300 min by means of gas proportional counting result in a detection limit of 0.1 TU at the University of Miami Tritium Laboratory [*Ostlund and Dorsey*, 1977]. Longer counting times (≤1000 min) can be used for smaller samples.

Researchers recently analyzed tritium using the helium 3 "in-growth" method [*Schlosser et al.*, 1989; *Solomon and Sudicky*, 1991). Tritium decays to ^{3}He. Pore water from the unsaturated zone is degassed of all He, sealed, and stored to decay to ^{3}He, allowing much higher precision and lower detection limits than do standard counting techniques. For example, a 20-mL water sample that is allowed to decay for 6 months would result in a detection limit of ~0.2 TU (R. Poreda, personal communication, 1995). The ^{3}He in-growth method for analyzing ^{3}H in unsaturated pore water samples should be distinguished from the ^{3}He in-growth dating method, which applies strictly to the saturated zone. Dating water using $^{3}H/^{3}He$ requires isolation of the ^{3}He from the atmosphere, which occurs only below the water table and provides the age of the water since it became isolated from the atmosphere [*Solomon et al.*, 1992]:

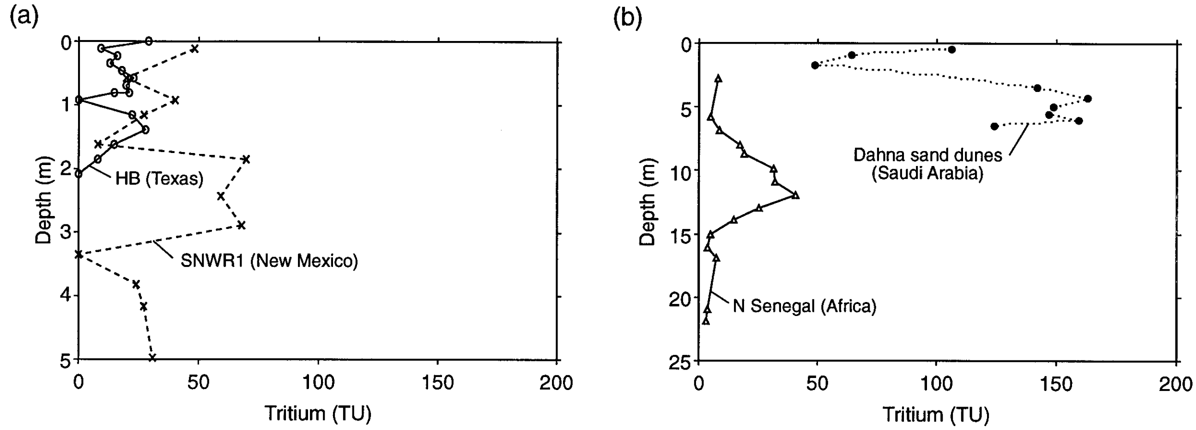

Figure 7. Profiles of ^{3}H concentrations (a) from the Chihuahuan Desert (Hueco Bolson (HB) [*Scanlon*, 1992]) and Sonoran Desert (SNWR1 [*Phillips et al.*, 1988]) and (b) from northern Senegal [*Aranyossy and Gaye*, 1992] (with permission from Gauthier-Villars Editeur) and Dahna sand dunes, Saudi Arabia (replotted from *Dincer et al.* [1974, Figure 11] with kind permission from Elsevier Science–NL, Amsterdam, Netherlands).

$$t_{^3\mathrm{H}/^3\mathrm{He}} = \lambda^{-1} \ln \left(\frac{^3\mathrm{He}}{^3\mathrm{H}} + 1 \right) \tag{12}$$

where $t_{^3\mathrm{H}/^3\mathrm{He}}$ is the ^{3}H/^{3}He age and λ is the ^{3}H decay constant.

At many arid sites, although the tritium bomb pulse within the root zone provides evidence of very low water fluxes [*Phillips et al.*, 1988; *Scanlon*, 1992a], accurate estimates of deep percolation below the root zone cannot be obtained from these data (Figure 7a). Deep penetration of the bomb pulse has also been found in sandy soils in arid settings [*Dincer et al.*, 1974; *Aranyossy and Gaye*, 1992] (Figure 7b). Comparison of ^{3}H and ^{36}Cl data from some arid sites showed deeper penetration of ^{3}H relative to ^{36}Cl, results that were attributed to enhanced downward movement of ^{3}H in the vapor phase [*Phillips et al.*, 1988; *Scanlon and Milly*, 1994]. Diffusion of ^{3}H in the vapor phase is limited if equilibration between liquid and gas phases occurs because the concentration of ^{3}H in the vapor phase is 5 orders of magnitude less than in the liquid phase, reflecting the different densities of water molecules in the two phases [*Smiles et al.*, 1995]. The liquid phase, in this case, acts as a large sink for tritium.

The method used to estimate water flux from bomb pulse tracer distributions is based on an assumption of steady downward advective flux, implying that the penetration depth of ^{36}Cl and ^{3}H increases linearly with time. Recent analytical studies by *Milly* [1996] suggest that the shallow distribution of these bomb pulse tracers can be attributed to episodic downward liquid flow and seasonal temperature gradients without invoking any mean vertical downward or upward water flux. The presence of ^{36}Cl and ^{3}H near the surface indicates little or no water flux below the root zone. High ^{3}H values (e.g., 1100 TU at 24-m depth, $\leq$ 162 TU at 109-m depth) have been found adjacent to the Beatty site, Nevada, that cannot readily be explained by liquid or combined liquid and vapor transport [*Prudic and Striegl*, 1995; *Striegl et al.*, 1996]. Because disposal practices at Beatty varied in the past and included disposal of as much as ~2000 m^3 of liquid waste, further research in ^{3}H movement at Beatty is warranted.

In some locations, bomb pulse ^{3}H has been found at depths greater than those initially expected. For example, bomb pulse ^{3}H was found as deep as ~450 m (105 TU; UZ-16 borehole) in Yucca Mountain (I. C. Yang, personal communication, 1995) and ~12 m (8.4 TU, RT18 borehole) in the Negev Desert [*Nativ et al.*, 1995]. These depths of ^{3}H migration, much greater than predicted by chloride mass balance data at these sites, may be attributed to preferential flow along fractures.

4. DIRECTION AND RATE OF WATER MOVEMENT

Although direction of water movement is a basic issue, it is not easily resolved at some sites, primarily because the water fluxes under consideration have a magnitude close to the errors inherent in measuring or in calculating these water fluxes. Second, a variety of driving forces in water movement may be important in arid settings, including water potential, gravitational potential, pneumatic potential, osmotic potential, and temperature. Third, the direction of water flux is likely to be spatially and temporally variable.

In this section we examine the various driving forces that can control the direction of water movement. Sediment heterogeneity also affects the direction of flow and is discussed later.

4.1. Liquid Flux

An initial examination of the simple system in which liquid flow is dominant shows that liquid water flux q_l is described by Darcy's law under steady flow conditions

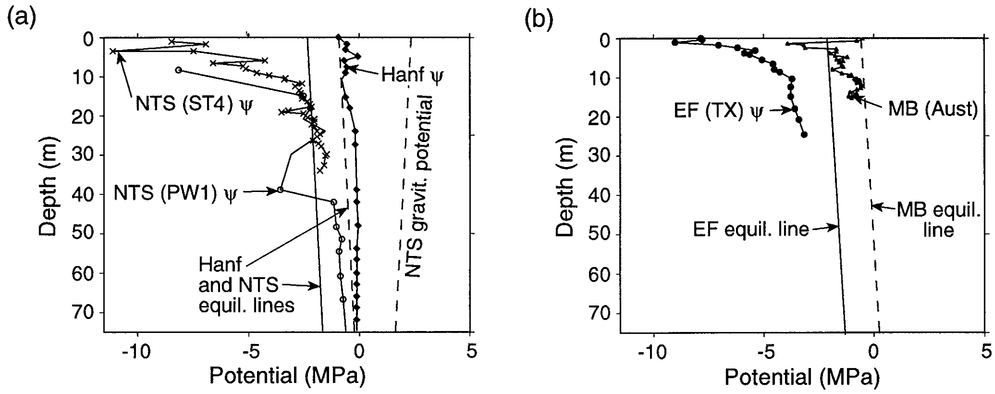

Figure 8. Evaluation of the direction of water movement according to the relationship between water potential profiles and the equilibrium line. Data are (a) from Hanford, Washington (Hanf; data from *Brownell et al.* [1975] as plotted by *Gee and Heller* [1985, Figure 4]), and Nevada Test Site, Nevada (NTS; profiles ST4 (shallow) and PW1 (deep) [*Estrella et al.*, 1993]), and (b) from Eagle Flat, Texas (EF111 [*Scanlon et al.*, 1997b]), and Murray Basin, South Australia (MB [*Jolly et al.*, 1989]). Equilibrium line refers to equilibrium matric potential that balances gravitational potential (Nevada Test Site data shown as an example).

according to equation (6). Evaluation of flow direction requires information on the hydraulic head (sum of matric and gravitational potential heads) gradient. Because matric potentials in natural interfluvial settings in arid systems are generally low, tensiometers cannot be used and thermocouple psychrometers are required that measure water potential (sum of matric and osmotic potential; see Tables 1 and 2 and Figure 3). The osmotic component of the water potential is generally negligible because zones where the magnitude of the osmotic potential is high in near-surface sediments generally correspond to zones where the magnitude of the water potential is also high [*Scanlon*, 1994]. Except in the shallow subsurface after rainfall, water (pressure) potentials measured in interfluvial settings in desert soils generally decrease (become more negative) toward the surface [*Jolly et al.*, 1989; *Fischer*, 1992; *Detty et al.*, 1993; *Scanlon*, 1994]. This upward decrease in water potentials suggests an upward driving force for liquid water flow.

One can also estimate the direction of water flow under steady flow conditions by comparing the measured matric or water potentials with the equilibrium matric potentials (Figure 8). If the vertical space coordinate z is taken as positive upward and zero at the water table, the equilibrium matric potential heads are the negative of the gravitational potential heads because matric and gravitational potential heads are balanced under static equilibrium (no flow) and their sum is a constant (0 in this case) (Figure 8). Under steady flow conditions, matric potentials that plot to the right of the equilibrium matric potential line indicate downward flow, and matric potentials that plot to the left of the equilibrium line indicate upward flow. At a site in Hanford, Washington, *Brownell et al.* [1975] (Figure 8a) found that measured water (pressure) potentials (approximately equal to matric potentials) plot to the right of the equilibrium line, indicating drainage. At several sites in Australia and in the southwestern United States, water (pressure) potentials plot to the right of the equilibrium line, indicating net upward water movement [*Jolly et al.*, 1989; *Fischer*, 1992; *Estrella et al.* 1993; *Scanlon*, 1994] (Figure 8b). At the Nevada Test Site this zone of net upward water movement is restricted to the upper 20–40 m (Figure 8a) [*Detty et al.*, 1993; *Sully et al.*, 1994]. Below 20–40 m, water potentials plot to the right of the equilibrium line, suggesting that liquid water at depth may be draining at this site.

4.2. Vapor Flux

Under dry conditions characteristic of arid settings, vapor flow may be significant. If the air phase is assumed to be static, vapor flux q_ν is given by

$$q_\nu = q_{\mathrm{I}\nu} + q_{\mathrm{T}\nu} = -D_{\mathrm{I}\nu}\nabla h - D_{\mathrm{T}\nu}\nabla T \tag{13}$$

where $q_{\mathrm{I}\nu}$ is the isothermal vapor flux, $q_{\mathrm{T}\nu}$ is the thermal vapor flux, $D_{\mathrm{I}\nu}$ is the isothermal vapor diffusivity, $D_{\mathrm{T}\nu}$ is the thermal vapor diffusivity, h is matric (pressure) potential head, and T is temperature. Isothermal vapor flux is driven by the matric (pressure) potential gradient and is unaffected by the temperature gradient, in a way similar to that of the liquid flux. Thermal vapor flux is driven by the temperature gradient and is unaffected by the matric potential gradient. Thermal vapor flux, resulting from variations in saturated vapor pressure according to temperature, is generally considered much more important than isothermal vapor flux. A temperature difference of 1°C at 20°C results in a greater difference in vapor density (1.04×10^{-3} kg m^{-3}) than does a 1.5-MPa difference in matric potentials from -0.01 MPa to -1.5 MPa (0.17×10^{-3} kg m^{-3}) [*Hanks*, 1992, p. 95]. The effects of temperature enter directly through temperature gradients and indirectly through temperature dependence of the matric (pressure) potential, hydraulic conductivity, and vapor diffusivity [*Scanlon and Milly*, 1994]. Thermally driven liquid flow is generally negligible under the low water contents characteristic of interfluvial arid settings [*Milly*, 1996].

Seasonal reversals in temperature gradients from upward movement in the winter to downward movement in the summer in the 2- to 12-m zone result in a net downward thermal vapor flux [*Fischer*, 1992; *Scanlon*, 1994] (Figure 1). Net downward thermal vapor fluxes are attributed to higher thermal vapor diffusivities as a result of higher temperatures in the summer when the gradients are downward. Below the zone of seasonal temperature fluctuations, the upward geothermal gradient provides an upward driving force for thermal vapor movement (Figure 1). Estimated values of local geothermal gradients are 0.06°C m^{-1} (Beatty site [*Prudic*, 1994]), 0.013°C m^{-1} (Nevada Test Site [*Tyler et al.*, 1996]), and 0.046°C m^{-1} (Hanford [*Enfield et al.*, 1973]). Calculated upward thermal vapor fluxes resulting from the upward geothermal gradient range from 0.02 mm yr^{-1} at the Nevada Test Site [*Sully et al.*, 1994] to 0.04 mm yr^{-1} at the Hanford site [*Enfield et al.*, 1973].

So far in our analysis we have considered vapor diffusion resulting from water (pressure) potential and temperature gradients only, but volatile contaminants may also diffuse as a result of concentration gradients. In addition to diffusion, advection may occur in the gas phase. Factors resulting in advective transport include barometric pressure fluctuations, density, wind, and temperature. In homogeneous, permeable media, *Buckingham* [1904] showed that the effect of barometric pressure fluctuations was small in relation to that of molecular diffusion. In fractured, permeable media, advective fluxes resulting from barometric pressure fluctuations may be orders of magnitude greater than diffusive fluxes and could result in upward movement of contaminated gases into the atmosphere [*Nilson et al.*, 1991]. A gas tracer experiment described by *Nilson et al.* [1992] confirms the importance of barometric pumping in causing upward gas movement in fractured tuff from a spherical cavity (depth ~300 m) created by underground nuclear tests at the Nevada Test Site. In areas of steep topography such as at Yucca Mountain, temperature- and density-driven topographic effects result in continuous exhalation of air through open boreholes at the mountain crest in the winter, as cold dry air from the flanks of the mountain replaces warm moist air within the rock-borehole system [*Weeks*, 1987]. Wind also results in air discharge from the boreholes that is ~60% of that resulting from temperature-induced density differences [*Weeks*, 1993]. Open boreholes greatly enhance the advective air flow at this site; numerical simulations indicate that water fluxes resulting from advective air flow under natural conditions (0.04 mm yr^{-1}) are 5 orders of magnitude less than those found in the borehole [*Kipp*, 1987] and similar in magnitude to estimated vapor fluxes as a result of the geothermal gradient (0.025 to 0.05 mm yr^{-1} [*Montazer et al.*, 1985]). These processes could cause drying of fractured rock uplands and could expedite the release of gases to the atmosphere [*Weeks*, 1993].

4.3. Water Flux

Water includes liquid and vapor phases. Analysis of data from several sites in the southwestern United States indicates that net water flux often occurs upward in the upper 20- to 40-m section of the unsaturated zone because water potentials plot to the left of the equilibrium line and total potential (matric [pressure] + gravitational) gradients are upward (Figure 8). In the zone of seasonal temperature fluctuations (2–12 m deep), upward liquid and isothermal vapor fluxes exceed downward thermal vapor fluxes (Figure 1). Upward water potential and temperature gradients at greater depths result in upward liquid and vapor fluxes (Figure 1).

Below the 20- to 40-m section, water potentials at the Nevada Test Site plot to the right of the equilibrium line [*Detty et al.*, 1993], indicating downward liquid and isothermal vapor flux under steady flow conditions, and upward thermal vapor flux due to the geothermal gradient (Figure 1). At this site, the upward thermal vapor flux (0.02 mm yr^{-1}), almost balanced by the downward liquid flux (0.03 mm yr^{-1}), results in a statistically insignificant net downward water flux of 0.01 mm yr^{-1} [*Sully et al.*, 1994]. At the Hanford site the upward thermal vapor flux (0.04 mm yr^{-1}), less than the downward liquid flux (0.30 mm yr^{-1}), results in a net downward water flux of 0.26 mm yr^{-1}. The larger flux at the Hanford site is attributed to higher water potentials (−0.1 MPa) relative to those at the Nevada Test Site (−0.6 MPa) [*Sully et al.*, 1994]. In the upper part of the unsaturated zone, different directions of liquid and vapor fluxes can therefore be important for evaluation of the transport of volatile and nonvolatile substances.

5. HOW IMPORTANT IS PREFERENTIAL FLOW?

Traditionally, piston-like flow, implying displacement of initial water by infiltrating water, was thought to be the dominant flow mechanism in the unsaturated zone. In the strict sense, piston flow refers to uniform displacement of solute or water without any mixing. True piston flow never occurs because of mixing due to molecular diffusion and microscopic water velocity variations. We therefore use the term "piston-like flow" instead of "piston flow" to represent predominantly matrix flow, or uniform flow, through the unsaturated matrix, in contrast to preferential flow, which bypasses much of the unsaturated zone. Data from many arid sites, particularly interfluvial settings that have unconsolidated sediments, suggest predominantly piston-like flow. Differences in velocities of solute (V_s) and wetting fronts (V_{wf}) in South Australia after vegetation clearing (Figure 9) could be predicted by the following equation, which assumes piston flow:

$$\frac{V_s}{V_{wf}} = \frac{\theta_f - \theta_i}{\theta_f} \qquad (14)$$

where θ_f is the final water content and θ_i is the initial water content [*Jolly et al.*, 1989]. Similar results were found in large field tracer experiments conducted in Las Cruces, New Mexico [*Young et al.*, 1992]. In these experiments the lag between the solute and the wetting front increased with depth, consistent with piston-like displacement of original pore water. Increases in initial water content resulted in increased lag between solute and wetting fronts. Single peaks in bomb pulse tracer distributions such as ^{36}Cl at sites in the Chihuahuan Desert site [*Scanlon*, 1992a], the Nevada Test Site [*Norris et al.*, 1987], and the Sonoran Desert site [*Phillips et al.*, 1988], are also consistent with piston-like flow (Figure 6). Although the aforementioned data suggest predominantly piston-like flow, they are not sensitive to small-scale preferential flow.

Preferential flow has received more emphasis in recent studies. With preferential flow the cross-sectional area of flow is reduced, and water bypasses much of the unsaturated medium, leading to corresponding increases in velocity and reduced sorption. Preferential flow was generally considered to become damped with depth; however, more recent studies suggest that this is not always true. Preferential flow can be divided into funneled flow, unstable flow, and macropore flow [*Steenhuis et al.*, 1994]. These three types of preferential flow are not mutually exclusive because unstable flow can occur in macropores (as will be described later). Funneled flow, occurring at textural interfaces, was extensively documented in glacial outwash deposits in Wisconsin under unsaturated conditions [*Kung*, 1990a, b; *Miyazaki*, 1993]. Laboratory experiments showed that when water application rates were ≤2% of the saturated hydraulic conductivity of the finer material, water flowed along the surface of the coarser layer [*Kung*, 1993]. Although funneled flow has not been found in arid settings, lateral flow in geologically layered materials resembles funneled flow where inclined beds and natural capillary barriers result in lateral flow. Such lateral flow along geologically layered materials has been hypothesized for Yucca Mountain, Nevada, on the basis of analytical solutions and numerical simulations [*Ross*, 1990; *Oldenburg and Pruess*, 1993].

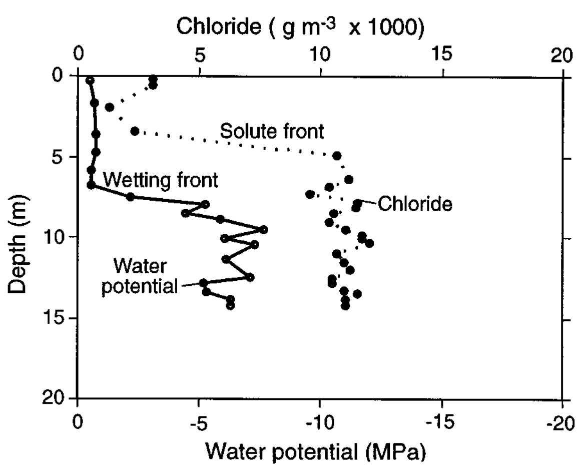

Figure 9. Piston-like flow evidenced by the lag between the wetting front and the solute front (modified from *Jolly et al.* [1989, Figure 2] with kind permission from Elsevier Science–NL, Amsterdam, Netherlands). The zone from the soil surface to the solute front represents water that infiltrated after the vegetation was cleared, the zone between the solute and wetting front represents displaced "preclearing" water, and the zone below the wetting front represents initial "preclearing" water.

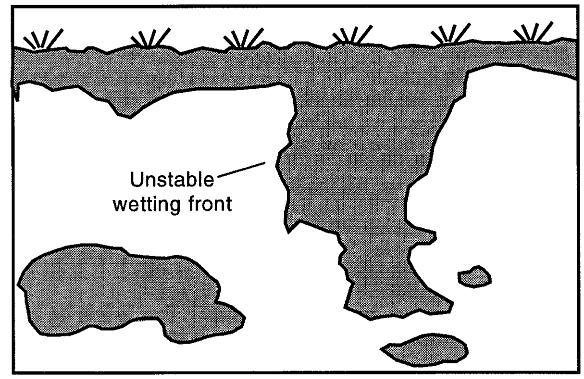

Figure 10. Example of unstable flow in water repellent soils after rainfall in the Netherlands (modified from *Hendrickx and Dekker* [1991]).

Unstable wetting fronts have been found in several field sites [*Starr et al.*, 1978, 1986; *Glass et al.*, 1988; *Hendrickx and Dekker*, 1991; *Selker et al.*, 1992; *Hendrickx et al.*, 1993] (Figure 10). *Chen et al.* [1995] provided an overview of instability and fingering in porous and fractured media. Important factors in the development of unstable flow in porous media include layering of sediment [*Hillel and Baker*, 1988; *Glass et al.*, 1989b], air entrapment [*Glass et al.*, 1990], and water repellency [*Hendrickx and Dekker*, 1991; *Ritsema et al.*, 1993; *Dekker and Ritsema*, 1994]. The absence of unstable wetting fronts in dune sands in an arid region of New Mexico led *Yao and Hendrickx* [1996] to evaluate conditions required for wetting-front instability. Many of the studies document that unstable wetting fronts were found in sandy, water-repellent soils because water repellency always results in unstable flow [*Hendrickx and Dekker*, 1991; *Ritsema et al.*, 1993; *Dekker and Ritsema*, 1994; *Ritsema and Dekker*, 1995]. Water tables are also shallow at many of these sites (0.5–1.5 m [*Ritsema et al.*, 1993]). *Hendrickx and Yao* [1996] subdivided infiltration rates into three regimes: low, medium, and high. Gravity-driven instabilities do not occur under low infiltration rates, where capillary and adsorptive forces are much greater than gravitational forces. Under high infiltration rates, wetting fronts remain stable if the infiltration rate approximates field-saturated hydraulic conductivity. Un-

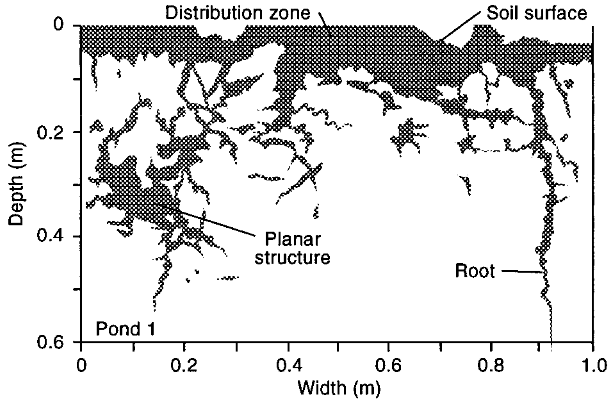

Figure 11. Example of preferential flow along roots as shown by FD&C dye [*Scanlon et al.*, 1997a].

der medium infiltration rates, stable wetting fronts are found when the total amount of infiltrating water is less than the amount of water required to wet a surface distribution layer. Additionally, the distribution layer has stable flow, and the thickness of this layer can be predicted by the same equation used by *Glass et al.* [1989a] to predict finger diameter [*Hendrickx and Yao*, 1996]. Application of these criteria to dune sands in New Mexico showed that all 2- and 10-year, and some 100-year return interval precipitation events were in the stable flow regime [*Hendrickx and Yao*, 1996]. Thus precipitation records and information on water repellency, water retention, and hydraulic conductivity of sediments at a site can be used to evaluate the potential for unstable flow.

Macropore flow refers to flow along noncapillary-size openings such as fractures, cracks, and root tubules (Figure 11). Important factors in evaluating macropore flow include sediment texture and structure and boundary conditions [*Flury et al.*, 1994]. Previous studies have shown that macropore flow is much greater in structured, fine-grained sediments than in structureless coarse-grained sediments [*Steenhuis and Parlange*, 1991; *Flury et al.*, 1994]. Whereas finger and funneled flow are eliminated under saturated conditions, it was previously thought that ponded conditions were required for macropore flow. Water ponds episodically in playas (ephemeral lakes) in arid systems. Detailed studies of playas have been conducted in the Southern High Plains of Texas, and preferential flow is inferred from the multipeaked character of a ^{3}H profile beneath a playa [*Scanlon et al.*, 1997a; *Scanlon and Goldsmith*, 1997] (Figure 12). Although ponding greatly enhances the potential for flow along macropores, such flow occurs under natural rainfall and sprinkler conditions also. Because water flow in noncapillary size pores occurs only when saturation is approached, macropore flow has been found mostly in humid sites that have higher precipitation [*Gish and Shirmohammadi*, 1991] or in arid settings subjected to ponding.

Much of the evidence for macropore flow in arid settings has been restricted to fractured media, such as tension fractures beneath fissured sediments in the Chihuahuan Desert [*Scanlon*, 1992b], fractured tuff in Yucca Mountain [*Fabryka-Martin et al.*, 1993], and fractured chalk in the Negev Desert [*Nativ et al.*, 1995]. Many fracture studies are based on laboratory experiments [*Nicholl et al.*, 1994]. *Glass et al.* [1995] proposed a "thought" experiment that may explain how preferential flow along fractures could transmit water over long distances, as seen at the Nevada Test Site [*Russell et al.*, 1987] and Yucca Mountain [*Fabryka-Martin et al.*, 1993; *Liu et al.*, 1995]. According to their thought experiment, gravity-driven fingers in inclined fractures are expected to persist over time. These fingers originate from point connections with water sources either at the surface or at a depth where perched zones occur. Water flow in the fractures should be only negligibly affected by water moving from the fracture into the matrix because of (1) the reduced flow area within the fractures (due to fingering and air entrapment), (2) reduced matrix storage capacity (most fractured rocks at Yucca Mountain are at or near satiated water content, i.e., near saturated with some entrapped air, at depth while still at low matric potential), and (3) vertical capillary barriers provided by surrounding fractures within the network that reduce conduction of water from one matrix block to another. With depth, fingers are expected to focus into a smaller number of stronger flow paths at the contact of larger aperture fractures, a concept contrary to prevailing ideas that preferential flow dissipates at depth when water moves into the matrix.

The continuity of preferred pathways, critical in macropore flow, depends on pathway type. Rock fractures can extend to great depths, whereas desiccation cracks and root tubules are generally fairly shallow. Although macropores are generally thought to provide pathways for enhanced downward liquid flow, macropores also provide pathways for gas and vapor move-

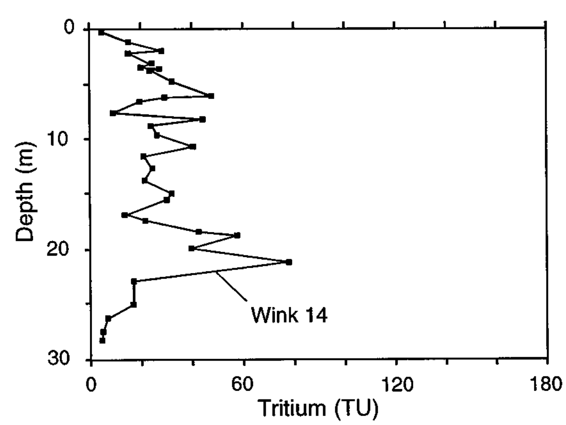

Figure 12. A deep multipeaked ^{3}H profile beneath a clay rich playa, indicating preferential flow (Wink 14 [*Scanlon et al.*, 1997a]).

ment and may enhance upward movement of volatile contaminants, as was suggested by *Weeks* [1993].

The type of contaminant helps determine the significance of preferential flow. Preferential flow is much more important for contaminants that exceed health standards in the parts-per-billion range, such as pesticides, than for contaminants that exceed health standards in the parts-per-million range, such as nitrate [*Steenhuis and Parlange*, 1991]. Nitrate contamination requires movement of the bulk of the pore water, which is much greater than the generally smaller water volume transported along preferred pathways. Arrival of the first 1% of the chemical at the groundwater is more readily accommodated by preferential flow than is the transport of the bulk of the mass.

Many of the studies evaluating preferential flow were either conducted in humid sites or performed on the basis of laboratory studies or theoretical analysis. Field evidence of preferential flow in arid settings has been found mostly in fractured rocks [*Fabryka-Martin et al.*, 1993; *Liu et al.*, 1995; *Nativ et al.*, 1995] and in fissured sediments [*Scanlon*, 1992b]. Bomb pulse tritium found at depths of ~12 m in an arid region in South Australia was attributed to preferential flow along the annular regions of eucalyptus roots [*Allison and Hughes*, 1983]. Few field studies show evidence of preferential flow at great depths in porous media in interfluvial arid settings, which may reflect (1) the absence of preferential flow in these settings, (2) the limited ability of various techniques to detect preferential flow in deep vadose zones, or (3) the difficulties of intercepting vertical preferred pathways by means of vertical boreholes.

6. CONTROLS ON WATER MOVEMENT

Water fluxes in arid regions have been shown to range widely both within and between various regions (Table 4). We can evaluate controls on unsaturated flow on the basis of comparisons of results from these studies. Primary controls such as pressure and temperature are discussed in the section on the direction of water movement. In this section we evaluate controls such as vegetation, climate, texture, and topographic setting.

6.1. Vegetation

Vegetation may be the most important control on water movement in desert soils. Uniformity in chloride profiles throughout the arid regions of the southwestern United States was attributed by *Phillips* [1994] to the ability of desert vegetation to control water fluxes regionally. Although annual precipitation and soil type varied widely among the sites examined by *Phillips* [1994], the chloride profiles were remarkably uniform. Because vegetation in arid regions is opportunistic, when the water application rate is increased, plant growth increases to use up the excess water. The opportunistic nature of desert vegetation is shown by higher concentrations of vegetation in areas of increased water flux, such as in ephemeral streams and in fissured sediments. When water supply is limited, plant activity decreases until water supply rates increase. Field studies have shown the importance of vegetation on local scales. Lysimeter studies in Hanford, Washington, and Las Cruces, New Mexico, showed deep drainage ranging from 10 to >50% of the annual precipitation in bare, sandy soils [*Gee et al.*, 1994]. The presence of plants at other sites at Hanford also greatly reduced deep drainage. Studies in Cyprus have found highest recharge rates in areas of sparse vegetation and lowest recharge rates in areas of bush vegetation [*Edmunds et al.*, 1988] (Table 4). Influence of vegetation is most clearly seen in areas where the vegetation cover has changed. In Australia, replacement of native mallee vegetation (deep-rooted eucalyptus trees) with crops resulted in an increase in recharge rates of at least an order of magnitude (from 0.1–0.9 mm yr^{-1} for native mallee regions to 4–28 mm yr^{-1} for pasture regions [*Cook et al.*, 1994]). Types of vegetation differ in how effectively they transpire water. In Hanford, Washington, water fluxes estimated in a grass site were much greater than water fluxes in nearby areas that had shrub vegetation [*Prych*, 1995]. Natural wildfires had resulted in replacement of shrubs with grass at this site. The effectiveness of vegetation in removing water from the subsurface was demonstrated in a lysimeter study at the Hanford site, where a lysimeter that had been bare for 3 years accumulated 150 mm of water in storage. The lysimeter subsequently became vegetated by deep-rooted plants (Russian thistle) that removed the excess water within a 3-month period to a depth of ~3 m [*Gee et al.*, 1994].

6.2. Climate and Paleoclimate

Although average annual precipitation is used to assess the potential for unsaturated flow, it is generally not a very good indicator of the rate of water movement. Data from various settings show little or no relationship between annual precipitation and water flux (Table 4). Seasonal distribution in precipitation is a better indicator of water flux in desert soils than is mean annual precipitation. Winter precipitation percolates through the soil more effectively than summer precipitation because evapotranspiration is low in the winter and the nature of precipitation varies seasonally. Winter precipitation in many parts of the world results from low-intensity, long-duration, frontal storms that are more likely to infiltrate in contrast to summer precipitation, which results from high-intensity, short-duration, convective storms. Snowmelt in the winter in some areas remains on the land surface longer, too, and infiltrates more readily.

As long-term mean annual precipitation rate decreases, variability in annual precipitation generally increases, and desert sites may experience many years of below-average precipitation followed by 1 or 2 years of normal or above-average precipitation. Because deep

TABLE 4. Water Fluxes in Various Arid Settings Throughout the World Estimated on the Basis of Different Measurement Techniques

Location	*Authors*	*Precipitation, mm yr^{-1}*	*Method*	*Water Flux, mm yr^{-1}*	*Topography/Texture/Vegetation*
S. Australia	*Allison et al.* [1985]	~300	chloride	>60	sinkholes
S. Australia	*Allison et al.* [1985]	~300	chloride	0.06–0.17	vegetated sand dunes
S. Australia	*Cook et al.* [1994]	260	chloride	0.1	sands, native vegetation
			chlorine 36	0.9	
S. Australia	*Cook et al.* [1994]	340	chloride	4–28	sand dunes, cleared vegetation
			chlorine 36	2–11	
			tritium	8–17	
Saudi Arabia	*Dincer et al.* [1974]	80	tritium	23	sand dunes
N. Senegal	*Aranyossy and Gaye* [1992]	395	tritium	22–26	sand dunes
Sudan	*Edmunds et al.* [1988]	225	chloride	0.25–1.28	interfluve sandy clay
Cyprus	*Edmunds et al.* [1988]	406	chloride	33–94	Fine-grained sands, sparse vegetation
			tritium	22–75	
			chloride	10	Fine-grained sands, bush vegetation
Israel	*Nativ et al.* [1995]	200	tritium	16–66	fractured chalk
			bromide	30–110	
Hueco Bolson, Texas, U.S.A.	*Scanlon* [1991]	280	chloride	0.01–0.7	ephemeral stream, silt loam
			chlorine 36	1.4	
			tritium	7	
Southern High Plains, Texas, U.S.A.	*Wood and Sanford* [1995]	460	tritium	77	playa, clay underlain by sand
New Mexico, U.S.A.	*Phillips et al.* [1988]	200	chloride	1.5–2.5	sandy loam to sand
			chlorine 36	2.5–3	
			tritium	6.4–9.5	
New Mexico, U.S.A.	*Stephens and Knowlton* [1986]	200	Darcy's law (unit gradient)	7–37	sand loam to sand
New Mexico, U.S.A.	*Stone* [1984]	385	chloride	0.8	cover sand
				4.4	sand hills
				≥12	playa clay
Las Cruces, New Mexico, U.S.A.	*Gee et al.* [1994]	230	lysimeter	87	loamy fine sand and silty clay loam, bare
Beatty, Nevada, U.S.A.	*Prudic* [1994]		chloride	2 (>10 m depth)	coarse texture, creosote bush
Nevada Test Site, U.S.A.	*Detty et al.* [1993]	125	liquid flux	0.03	Darcy's law depth 75–180 m
			vapor flux	0.02	
			net flux	~0	
Nevada Test Site, U.S.A.	*Tyler et al.* [1992]	125	tritium	600	subsidence crater, coarse sediment
Yucca Wash, Nevada, U.S.A.	*Norris et al.* [1987]	170	chlorine 36	1.8	ephemeral stream
Ward Valley, California, U.S.A.	*Prudic* [1994]	117	chloride	0.03–0.05 (>10 meter depth)	alluvial fan
Hanford, Washington, U.S.A.	*Prych* [1995]	160	chloride	0.01–0.3	shrub, sand
			chloride	0.4–2.0	grass, sand
			chlorine 36	5.1	grass, sand

percolation may occur only in the years of above-average rainfall, desert soils may be characterized by episodic flow. Although many researchers report water fluxes annually, for general purposes of comparing different techniques or for convenience, this method of reporting fluxes may be unrealistic. Long-term monitoring of physical parameters is required to evaluate episodic flow; however, such records are unavailable at most sites. Monitoring of water content in ~100 boreholes in Yucca Mountain from 1984 through 1993 showed that water content remained low during a 6-year drought but increased beginning in the winter of 1991 through 1993 as a result of increased precipitation [*Flint and Flint*, 1995]. Because monitoring of physical parameters represents only the monitoring period, evaluating how representative this time period is with respect to long-term climate is important for predictive purposes.

Distribution of environmental tracers has been used for evaluating water fluxes over a much longer timescale. Low chloride concentrations at depth in the southwestern United States (Figure 5) have been attributed to higher water fluxes during the Pleistocene, when the climate was cooler and wetter [*Scanlon*, 1991; *Phillips*, 1994; *Tyler et al*., 1996]. Higher water (pressure) potentials at depth in these arid regions may be attributed to drainage of older, Pleistocene water [*Scanlon*, 1994; *Tyler et al*., 1996]. Chloride and water potential data suggest that deep vadose zones in arid regions may reflect Pleistocene climate and that the shallower zone may have been drying since the Pleistocene. The deep vadose zone is therefore not in equilibrium with the current surface climate. Numerical simulations of long-term climate changes at Yucca Mountain suggest that the upper 75 m may have been undergoing long-term drying for the past 3000 years [*Flint et al*., 1993]. The cyclic climate inputs are damped with depth, and simulations suggest steady state conditions at depths ≥250 m.

Another factor of importance with respect to climate change and waste disposal is that sites that are now arid may not always be arid. A NAS panel evaluated the impact of climate change on high-level radioactive waste disposal at Yucca Mountain [*NRC*, 1995]. The Earth is currently in an interglacial phase. Although the Earth will probably not return to a glacial climate in the next few hundred years, the possibility cannot be ruled out [*NRC*, 1995]. A return to glacial conditions is probable within a 10,000-year time frame, which is the time required for high-level radioactive waste to be isolated from the accessible environment in the United States. A cooler, wetter climate associated with glacial times would result in increased water fluxes through the unsaturated zone. The ~300-m-thick unsaturated section overlying the proposed high-level radioactive waste disposal repository at Yucca Mountain would result in a large time lag of the order of hundreds to thousands of years between surface climate change and water fluxes at the level of the repository [*NRC*, 1995]. Climate changes of the order of hundreds of years would therefore be damped out at the depth of the proposed repository.

Although the time period required for isolation of low-level radioactive waste (1000 years) is much shorter than that required for high-level radioactive waste, low-level radioactive waste is buried at shallow depths; therefore the effects of damping of climate changes would be less for shallow burial sites. Environmental tracers such as chloride provide some indication of potential increases in water fluxes associated with glacial climates. A review of chloride profiles at several sites in the southwestern United States suggests that water fluxes would increase by a factor of ~20 [*Phillips*, 1994]. The highest water fluxes estimated during glacial times at these sites were ~3 mm yr^{-1}, which is still low. Thus the effect of climate change on water flux should be considered in siting the disposal facilities.

6.3. Sediment Texture

Texture of surficial sediments can greatly affect water movement in the unsaturated zone. Fine-grained surface soils provide a large storage capacity and retain infiltrated water near the surface, where it is available for evapotranspiration. As was discussed earlier, macropore flow is much more common in highly structured, fine-grained sediments [*Flury et al*., 1994; *Bronswijk et al*., 1995]. Coarse-grained sediments allow water to penetrate more deeply into the soil, commonly below the zone from which it can be evapotranspired. For example, the estimated water flux was high in a sand dune area in Saudi Arabia according to tritium data (23 mm yr^{-1} [*Dincer et al*., 1974]; see Figure 7b and Table 4), representing ~30% of the long-term mean annual precipitation (80 mm yr^{-1}). *Cook et al*. [1992] noted an apparent negative correlation between clay content in the upper 2 m and the recharge rate. The concept of fine-grained surficial sediments providing large storage capacities is also employed in engineered barrier design. At the Hanford site, the texture and thickness of the sediment in an engineered barrier were chosen to provide storage capacity sufficient for 3 times the long-term mean annual precipitation [*Wing and Gee*, 1994]. Thickness of surficial unconsolidated sediments on top of fractured rock is also an important control on water fluxes. At Yucca Mountain, water penetration and environmental tracer distribution indicated minimal water fluxes in areas of thick alluvial cover over fractured tuff [*Fabryka-Martin et al*., 1993]. Similarly, the thickness of loess on fractured chalk in the Negev Desert greatly reduced water fluxes through the chalk [*Nativ et al*., 1995].

Heterogeneity and layering of sediments are also important in controlling water movement. Textural heterogeneity occurs at a variety of scales; small-scale, local heterogeneity may not be very important in extremely dry sediments, typical of interfluvial settings in arid regions, because most water is adsorbed to grain surfaces, and much of the water flux may occur in the vapor phase. In areas of ponded surface water, however, small-scale variations in sediment texture may have a greater effect on flow.

Layering of sediments reduces water fluxes. Where fine-grained sediments overlie coarse-grained sediments, a capillary barrier is formed, and water will not flow into the coarse layer until the overlying fine layer is close to saturation. Where interfaces between the different layers are sloped, lateral flow can occur. Capillary barriers occur in the natural system at a variety of scales. Studies by *Kung* [1990a, b] indicate that sloping layers can result in unstable flow at the downstream end when sufficient water accumulates in the fine-grained material to flow into the underlying coarse material [*Steenhuis et al*., 1991]. One of the conceptual models developed for Yucca Mountain suggests that the layered nonwelded tuff units may act as capillary barriers beneath the welded fractured units [*Montazer and Wilson*, 1984]. The capillary barrier concept is also used in engineered-barrier design to maximize evapotranspiration, minimize deep percolation, and (where such layers are sloped) allow lateral drainage [*Wing and Gee*, 1994].

Where fine-grained layers underlie coarser layers, perched water conditions can occur. Numerical simulations indicate that for perching to occur, downward water flux should exceed saturated hydraulic conductivity of the perching layer by an order of magnitude [*Schneider and Luthin*, 1978]. Perched water has been found in the vadose zone at Yucca Mountain [*Burger and Scofield*, 1994] and beneath ephemeral lakes (playas) in the Southern High Plains [*Mullican et al*., 1994].

6.4. Topography

Topographic setting may also play an important role in controlling unsaturated flow. Measurement of physical parameters and environmental tracer distributions in various topographic settings at Yucca Mountain showed that water fluxes were highest in active channels where surface runoff occurs [*Flint and Flint*, 1995]. In South Australia, because sinkholes focus surface water, much higher water fluxes were found beneath sinkholes (≥60 mm yr^{-1}) than in surrounding vegetated topographic settings (0.06–0.17 mm yr^{-1} [*Allison et al*., 1985]; see Table 4). Ephemeral lakes or playas in the Southern High Plains of Texas and New Mexico also focus recharge, and estimated water fluxes range from ≥12 mm yr^{-1} [*Stone*, 1990] to 77 mm yr^{-1} [*Wood and Sanford*, 1995]. Fissured sediments in the Chihuahuan Desert of Texas concentrate surface runoff, and water fluxes are much higher beneath these fissures than in surrounding areas [*Scanlon*, 1992b]. Nuclear subsidence craters at the Nevada Test Site are also characterized by high water fluxes (~600 mm yr^{-1}) as evidenced by high tritium concentrations and high water (pressure) potentials relative to profiles 207 m from the crater center [*Tyler et al*., 1992].

These studies suggest that local zones of high water flux, typical of arid settings, are generally found in topographic depressions where surface water collects, such as washes, playas, excavations, and sinkholes. Whereas the total surface area occupied by these features may be extremely small (e.g., 2% in the case of active channels in Yucca Mountain [*Flint and Flint*, 1995]) high flows beneath these features may be critical for transporting contaminants rapidly. Use of areally averaged recharge rates to predict contaminant transport would greatly underestimate the transport rates in these areas.

Paleotopography may also have affected the response of different sites to wetter climatic conditions during previous glacial periods. Low chloride concentrations deeper than 10 m at a site in the Amargosa Desert, Nevada, are attributed to increased precipitation and more frequent flooding of the Amargosa River at this site [*Prudic*, 1994]. Studies at the Nevada Test Site show much higher water fluxes in an area where surface runoff concentrated from the surrounding mountains during previous glacial maxima [*Tyler et al*., 1996].

7. NUMERICAL MODELING

The complexity of flow in the shallow unsaturated zone of desert systems requires the use of numerical models to evaluate flow processes and to analyze interactions and feedback mechanisms between various controlling parameters. A variety of codes are available to simulate flow and transport. Simulation of flow in very dry unsaturated systems can be computationally difficult, however. Conservation of mass was a problem with traditional head-based codes, but it has been overcome with the mixed formulation of Richards' equation, which uses water content in the time derivative and head in the space derivative [*Celia et al*., 1990]. Large execution times were also a problem that has been reduced by transformations of Richards' equation [*Kirkland et al*., 1992; *Pan and Wierenga*, 1995]. Representation of water retention functions is also important for dry systems. Traditionally, residual water content was treated as a fitting parameter in water retention functions; however, resultant water contents were commonly greater than initial water contents in simulations in arid settings [*Hills and Wierenga*, 1994]. More realistic water retention functions have been developed recently that incorporate the full range of water content from saturation to air-dry conditions [*Milly and Eagleson*, 1982; *Rossi and Nimmo*, 1994; *Fayer and Simmons*, 1995].

The performance of various codes in simulating field-tracer experiments conducted in Las Cruces, New Mexico, was evaluated as part of the International Cooperative Project on Validation of Geosphere Transport Models (INTRAVAL), which represented an international study of validation of models for flow and transport. An extensive database characterized the hydraulic properties at this site and included ~600 measurements of bulk density, saturated hydraulic conductivity, and water retention. Two-dimensional models that assumed a heterogeneous porous medium performed no better than one-dimensional models that assumed a homogeneous porous medium [*Hills and Wierenga*, 1994]. The experiments at Las Cruces were conducted on bare soil and excluded evaporation. Detailed simulations of flow in a natural system require nonisothermal liquid and vapor flow, atmospheric forcing, and water uptake by roots. Very few codes incorporate all these features. Because simulation of preferential flow is extremely complicated, new codes need to be developed to address this issue. A code developed by *Nieber* [1996] successfully simulates unstable flow. Several investigators are simulating flow in fractured rock on the basis of data from the Yucca Mountain site, and some of these studies attempt to reproduce the tracer data that suggest preferential flow [*Wolfsberg and Turin*, 1996].

Previous studies that included numerical simulations provide valuable insights into unsaturated-flow processes. Simulations of flow in a bare soil show net downward thermal vapor flux in response to seasonal temperature gradients in the shallow subsurface [*Scanlon and Milly*, 1994]. Results of flow simulations of engineered barriers agree with field data from lysimeters at the Hanford site, Washington [*Fayer et al*., 1992]. This study shows that hysteresis is important in simulating breakthrough of capillary barriers. Numerical modeling of flow at Yucca Mountain evaluated the effect of long-term climatic change on net infiltration and showed that amplitude and frequency of climate change are important factors [*Flint et al.*, 1993]. Below 250 m, climatic changes having a frequency ≤50,000 yr were damped out.

Evaluation of potential sites for disposal of waste, such as low- and high-level radioactive waste, requires performance assessment to develop a quantitative understanding of system behavior. For high-level nuclear waste disposal in the United States, performance assessment is required for time periods of 10,000 years or more. Although performance assessment of many sites includes rigorous parameter uncertainty analysis, the main source of uncertainty generally results from conceptual model uncertainty. Performance assessment has used spatially and temporally invariant upper boundary conditions, even though the long time and space scales considered in performance assessment require the use of spatially and temporally varying upper boundary conditions that relate to topography and climate. If one is trying to predict future behavior of a 10,000-year time period, future climatic changes should be incorporated into the performance assessment. Whereas the U.S. Nuclear Regulatory Commission is promoting a probabilistic approach to performance assessment, a recent NAS panel on Yucca Mountain suggested that if compliance is met in bounding estimates that are based on upper or lower limits of parameters that result from conservative assumptions, more complex analysis is not needed [*NRC*, 1995]. This does not preclude performance monitoring to evaluate whether simplistic models of flow and transport are valid.

8. IMPLICATIONS FOR CONTAMINANT TRANSPORT RELATED TO WASTE DISPOSAL

The natural characteristics of a site are important for long-term (≥ decades) disposal of waste because the natural system is ultimately relied on to minimize waste migration. The attributes of the natural system are difficult to characterize, however, because of the low water fluxes and limitations of monitoring instruments, as discussed earlier. To overcome some of these problems, multiple independent lines of data are required to increase confidence in results. Despite the difficulties in characterization, a larger margin of error can be tolerated in arid settings than in humid settings because of the naturally low water fluxes in porous media in interfluvial arid settings in porous systems. Important attributes of the natural system include direction and rate of water movement and the spatial and temporal variability in water fluxes. The type of medium (porous or fractured) is very important because of the higher potential for preferential flow in fractured systems. The vegetative cover is also important because it removes much of the infiltrated water from the subsurface.

Engineered designs and disposal practices are critical for developing a reliable disposal system. Although much information exists on site characteristics, our knowledge of the performance of engineered systems is generally limited. Ideally, an engineered system should mimic the natural system as much as possible, and the performance of various design elements of engineered systems should be rigorously tested in arid regions. Detailed studies of a capillary barrier system are being conducted at Hanford, Washington [*Wing and Gee*, 1994]. Trench-cap demonstration units will also be constructed at Ward Valley, California, to evaluate the performance of these systems [*NRC*, 1995]. Past disposal practices have often greatly enhanced the likelihood of contamination at various sites. Disposal of liquid wastes at the Beatty site (~2000 m^3 between 1962 and 1975), for example, may have resulted in the large tritium concentrations found near that disposal site [*Striegl et al.*, 1996]. Future disposal practices of low-level radioactive waste will therefore be restricted to solid wastes. Restriction of waste to a solid form does not necessarily preclude contamination because water percolating through the unsaturated zone could dissolve the waste. Critical components of near-surface engineered systems include the vegetative cover to remove water by evapotranspiration, the storage capacity of surficial sediments to hold water in the shallow zone, where it can be readily evapotranspired, and biointrusion barriers to limit human, animal, and plant intrusion into the waste. Capillary barriers not only increase the storage capacity of surficial sediments but also serve to limit biointrusion.

Monitoring of these engineered systems will be important to ensure that they perform as designed and to provide data for performance assessment. Although monitoring of low-level radioactive waste disposal facilities is required for at least 30 years, the life span of many of the monitoring instruments, such as the thermocouple psychrometer, is much shorter than 30 years. Many systems are currently available for monitoring disposal facilities, such as the Science and Engineering Associates for the Membrane Instrumentation and Sampling Technique (SEAMIST) system, which consists of an impermeable membrane that is turned inside out (everted) under pressure and that can be used to pull various logging tools through tunnels below the waste or in the cover system [*Keller*, 1991]. This system has the advantage of being readily able to incorporate newly developed technologies.

9. IMPORTANT AREAS OF FUTURE RESEARCH

Our review suggests that although within the last couple of decades considerable progress has been made in our understanding of unsaturated-flow processes in arid regions, areas exist where future research should be directed. With respect to techniques that can be used to quantify unsaturated flow, additional research should be done to evaluate the effects of instrument installation on the monitoring data. Such research could include numerical simulations or laboratory or controlled field experiments to address this issue. Most techniques used to monitor the energy status of pore water are not very robust and have a limited life span. Because regulations for waste disposal, including low-level and high-level radioactive waste disposal, require monitoring for decades, efforts should be made to develop robust instrumentation that can be used for monitoring energy potentials over long time periods. Use of time domain reflectometry in arid regions is not very widespread now, but TDR is a promising tool for detailed monitoring near the land surface atmosphere boundary, and it will most likely provide valuable information on this critical boundary as well as integrate easily with remote-sensing studies. Because unsaturated hydraulic conductivity is the most uncertain parameter, considerable effort should be directed toward developing better techniques of quantifying or estimating this parameter. The applicability of traditional methods of estimating unsaturated hydraulic conductivity from capillary bundle models should be critically evaluated for arid systems where film flow may be dominant. Although noninvasive monitoring techniques have only recently been used in vadose zone studies, they should be the focus of future studies to quantify relationships between geophysical response and water fluxes in various settings.

Establishing the direction of water movement in arid settings is extremely difficult because of the complex interaction of forces. Because it has been 40 years since *Philip and de Vries* [1957] established the theoretical framework for liquid and vapor flow, it should be revisited in light of all the work that has been conducted since then. The importance of preferential flow in arid regions should be critically examined as well. Although field studies in a number of regions demonstrate preferential flow in fractured media, field studies of preferential flow in porous media in deep vadose zones in arid settings are extremely limited. The idea that macropore flow in shallow, unsaturated, porous media can be extrapolated to great depths in arid regions has not been shown in the field, nor has it been thoroughly studied. Likewise, indiscriminate extrapolation of results of preferential flow studies that have been conducted in humid regions that have shallow water tables should be avoided. The extent to which preferential flow persists or dissipates with depth is important in thick, unsaturated, layered systems. Techniques used to evaluate unsaturated flow should therefore be critically examined to ensure that the presence or absence of preferential flow is not simply an artifact of the measurement process. Preferential flow is an issue critical in the siting of waste disposal facilities in arid regions and in the evaluation of contaminant transport and remediation.

Vegetation may be the dominant control on water fluxes in arid settings, however, and various aspects of this issue should be examined from laboratory, field, and numerical modeling perspectives. The effect of climate and paleoclimate on water fluxes should also be intensively studied to help predict unsaturated flow thousands of years into the future, as required by the high-level radioactive waste disposal program at Yucca Mountain.

10. CONCLUSIONS

Much of the work in unsaturated-zone hydrology has been conducted in humid regions; however, fundamental differences between humid and arid regions restrict the applicability of results from humid sites to arid sites. In addition, a wider variety of techniques are required to quantify unsaturated flow in the much drier unsaturated systems in arid regions.

Many arid area studies suggest that using environmental tracers to quantify unsaturated flow is more appropriate than physical approaches because hydraulic conductivity can vary over orders of magnitude. Both approaches should be used, however, because physical data provide information on current processes, whereas environmental tracers provide information on longer term, net water fluxes. A variety of environmental tracers should also be used because some are restricted to liquid phase flow, whereas others are found in liquid and vapor phases. Noninvasive techniques, such as electromagnetic induction, should be further investigated, particularly for evaluation of contaminated sites. Multiple independent lines of data are required to increase confidence in conceptual models of flow and transport in arid regions.

Low water fluxes and inaccuracies in techniques for quantifying such fluxes make it difficult to resolve basic issues such as direction and rate of water movement. The direction of water movement is difficult to evaluate in many arid sites because unsaturated systems are commonly extremely dry and because water flows in liquid and vapor phases in response to water potential, gravitational potential, pneumatic potential, and temperature gradients that are temporally and spatially variable. Temporal variability in water flow occurs at a variety of scales, including diurnal, seasonal, decadal, and millennial intervals, all of which are commonly controlled by climate. Short-term climatic fluctuations are preserved in the shallow subsurface, whereas longer-term paleoclimatic fluctuations are preserved over the thick unsaturated sections found in many arid settings. At many sites, water fluxes were much higher during previous glacial periods.

Vegetation may be the most important control on unsaturated water movement, as is shown by high rates of water movement in areas of coarse, bare soil and by negligible water movement in vegetated areas. Surface topography also plays an important role in controlling water movement by focusing unsaturated flow in topographic depressions that pond frequently. Increasing the thickness of unconsolidated sediments on fractured media in arid regions greatly decreases unsaturated water fluxes, as is shown by studies at Yucca Mountain, Nevada, and the Negev Desert, Israel.

Field evidence of preferential flow in arid regions has generally been restricted to fractured media, as evidenced by deep penetration of bomb pulse tracers in fractured tuff at Yucca Mountain and in fractured chalk in the Negev Desert. Although many studies suggest predominantly piston-like flow in porous media in interfluvial arid settings, some of the techniques used may not be sensitive to small percentages of preferential flow. Recent studies suggest that unstable flow, which is driven by gravity, should be negligible in porous media in many arid regions because of the dominance of capillary and adsorptive forces over gravity forces in these areas. Evaluation of preferential flow is much more difficult in arid regions than in humid regions because the thickness of the unsaturated section is greater and short-term applied tracer experiments cannot be used in the thick vadose zones typical of many arid regions.

Because of (1) many uncertainties in determining water fluxes in arid areas, (2) extensive spatial and temporal variability in properties, (3) vegetation, and (4) precipitation, generalized conclusions about recharge rates at a specific site are difficult to make. Detailed investigations are required to determine the nature, magnitude, and direction of water fluxes at specific locations.

GLOSSARY

Advection: movement of solute with the flowing fluid; movement of gas in response to total pressure gradient.

Capillary barrier: layer of fine sediment underlain by layer of coarse sediment that restricts downward movement of water because of the difference in the size of the capillaries. Water enters the underlying coarse layer when the matric potential in the fine layer increases sufficiently to overcome the water entry potential of the coarse layer.

Diffuse flow: movement of water into the unsaturated zone over large areas, as opposed to focused or concentrated flow.

Diffusion: movement of a substance, such as solute or vapor, along a concentration gradient.

Electromagnetic induction: technique to measure apparent electrical conductivity by electromagnetically inducing currents in the ground. Under low values of induction number, the secondary magnetic field is a linear function of conductivity.

Funneled flow: form of preferential flow that occurs when textural interfaces cause lateral water flow and accumulation of water in low regions.

Gravitational potential: change in energy per unit volume of water associated with change in the position of a body in the Earth's gravitational field. The reference state is generally defined as the land surface or the water table. Gravitational potential energy decreases with depth.

Heat dissipation probe: device used to measure matric potential in the unsaturated zone on the basis of variation in the rate of dissipation of a thermal pulse with water content. The probe is calibrated at different matric potentials.

Hydraulic conductivity: ability of material to conduct water; proportionality constant between water flux and hydraulic head gradient in Darcy's law.

Hydraulic head: sum of matric and gravitational potential heads.

Infiltration: rate of water movement from the surface to the subsurface.

Lysimeter: device for measuring water loss from soil and plants into the atmosphere. There are nonweighable and weighable lysimeters. Nonweighable lysimeters measure water storage changes indirectly (i.e., with a neutron probe and from inflow-outflow analysis), whereas weighable lysimeters measure storage changes gravimetrically.

Macropore flow: form of preferential flow in which water flows along noncapillary-size openings such as fractures, cracks, and root tubules.

Matric potential: change in energy per unit volume of water that results from the attraction of water to the solid matrix material.

Neutron probe: instrument used to monitor water content in the unsaturated zone.

Percolation or drainage: penetration of water below the shallow subsurface, where most evapotranspiration occurs.

Performance assessment: evaluation of future performance of a system on the basis of a quantitative understanding of system processes. Performance assessment generally includes long-term numerical simulations of system performance that incorporate uncertainties in conceptual models and in system parameters.

Piston-like flow: uniform downward movement of water through the unsaturated zone that displaces existing water without bypassing it.

Pneumatic potential: energy per unit volume of water resulting from changes in air pressure.

Potential energy: energy resulting from position of a body in a force field, such as gravitational, capillary, and osmotic force fields. Differences in potential energy can be used to determine the direction of water movement under isothermal conditions because water flows from regions of high to regions of low total potential energy. Potential energy is generally expressed as energy per unit volume (joules per cubic meter, equivalent to pressure units of newtons per square meter or pascals).

Preferential flow: nonuniform downward water movement along preferred pathways that bypasses much of the matrix and includes funnel flow, unstable flow, and macropore flow.

Recharge: addition of water to the aquifer.

Solute potential: equivalent to osmotic potential, change in energy per unit volume of water associated with the addition of solutes to pure, free water.

Suction lysimeter: device used to extract pore water from unsaturated media for chemical analysis.

Thermocouple psychrometer: device that measures relative humidity of water vapor in the sediment or rock sample, which is related to water potential ψ (energy per unit volume) through the Kelvin equation (equation (3)).

Time domain reflectometry: technique used to measure water content in unsaturated material on the basis of variation in the dielectric constant of the material with water content.

Unsaturated zone: zone in which pore spaces contain both water and air.

Unstable flow or fingering: form of preferential flow used to describe downward water movement in columns that may result from sediment layering, air entrapment, or water repellency.

Vadose zone: zone between land surface and regional water table.

Water activity: thermodynamic activity of water; relative humidity.

Water activity meter: device that measures water activity (relative humidity) of water vapor in sediment or rock samples and which is related to water potential through the Kelvin equation (equation (3)).

Water content: amount of water in unsaturated media; can be expressed gravimetrically (mass of water per mass of dry unsaturated material) or volumetrically (volume of water per volume of unsaturated material).

Water flux: volume of water flowing per unit cross sectional area per unit of time.

Water potential: pressure potential, sum of matric and osmotic potentials, can be measured by thermocouple psychrometers or water activity meter.

Water retention function: relationship between matric potential and water content.

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